Hydrogeology of, and simulation of ground-water flow in, the Pohatcong Valley, Warren County, New Jersey

By Glen B. Carleton and Alison D. Gordon

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Contents

Figures

Tables

Conversion Factors, Datums, and Well-Numbering System

Inch/Pound to SI

Multiply	By	To obtain
	Length	
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
Area		
square foot (ft^2)	0.09290	square meter (m^2)
square mile (mi^2)	2.590	square kilometer (km^2)
	Volume	
gallon (gal)	3.785	liter (L)
gallon (gal)	0.003785	cubic meter (m^3)
million gallons (Mgal)	3,785	cubic meter (m^3)
cubic foot (ft^3)	0.02832	cubic meter (m^3)
	Flow rate	
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
cubic foot per second per square mile [(ft^3/s)/mi^2]	0.01093	cubic meter per second per square kilometer [(m^3/s)/km^2]
gallon per minute (gal/min)	0.06309	liter per second (L/s)
inch per year (in/yr)	25.4	millimeter per year (mm/yr)
	Mass	
pound, avoirdupois (lb)	0.4536	kilogram (kg)
Specific capacity		
gallon per minute per foot [(gal/min)/ft)]	0.2070	liter per second per meter [(L/s)/m]
Hydraulic conductivity		
foot per day (ft/d)	0.3048	meter per day (m/d)
	Transmissivity*	
foot squared per day (ft^2/d)	0.09290	meter squared per day (m^2/d)

Vertical coordinate information is referenced to the North American Vertical Datum of 1988 (NAVD 88).

Horizontal coordinate information is referenced to the North American Datum of 1983 (NAD 83).

Altitude, as used in this report, refers to distance above the vertical datum.

*Transmissivity: The standard unit for transmissivity is cubic foot per day per square foot times foot of aquifer thickness [(ft^3/d)/ft^2]ft. In this report, the mathematically reduced form, foot squared per day (ft^2/d), is used for convenience

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Abstract

A numerical ground-water-flow model was constructed to simulate ground-water flow in the Pohatcong Valley, including the area within the U.S. Environmental Protection Agency Pohatcong Valley Ground Water Contamination Site. The area is underlain by glacial till, alluvial sediments, and weathered and competent carbonate bedrock. The northwestern and southeastern valley boundaries are regional-scale thrust faults and ridges underlain by crystalline rocks. The unconsolidated sediments and weathered bedrock form a minor surficial aquifer and the carbonate rocks form a highly transmissive fractured-rock aquifer. Ground-water flow in the carbonate rocks is primarily downvalley towards the Delaware River, but the water discharges through the surficial aquifer to Pohatcong Creek under typical conditions.

The hydraulic characteristics of the carbonate-rock aquifer are highly heterogeneous. Horizontal hydraulic conductivities span nearly five orders of magnitude, from 0.5 feet per day (ft/d) to 1,800 ft/d. The maximum transmissivity calculated is 37,000 feet squared per day. The horizontal hydraulic conductivities calculated from aquifer tests using public supply wells open to the Leithsville Formation and Allentown Dolomite are 34 ft/d (effective hydraulic conductivity) and 85 to 190 ft/d (minimum and maximum hydraulic conductivity, respectively, yielding a horizontal anisotropy ratio of 0.46). Stream base-flow data were used to estimate the net gain (or loss) for selected reaches on Brass Castle Creek, Shabbecong Creek, three smaller tributaries to Pohatcong Creek, and for five reaches on Pohatcong Creek. Estimated mean annual base flows for Brass Castle Creek, Pohatcong Creek at New Village, and Pohatcong Creek at Carpentersville (from correlations of partial- and continuous-record stations) are 2.4, 25, and 45 cubic feet per second (ft^3/s) (10, 10, and 11 inches per year (in/yr)), respectively.

Ground-water ages estimated using sulfur hexafluoride (SF_6), chlorofluorocarbon (CFC), and tritium-helium age-dating techniques range from 0 to 27 years, with a median age of 6 years. Land-surface and ground-water water budgets were calculated, yielding an estimated rate of direct recharge to the surficial aquifer of about 23 in/yr, and an estimated net recharge to the ground-water system within the area underlain by carbonate rock (11.4 mi^2) of 29 in/yr (10 in/yr over the entire 33.3 mi^2 basin).

A finite-difference, numerical model was developed to simulate ground-water flow in the Pohatcong Valley. The four-layer model encompasses the entire carbonate-rock part of the valley. The carbonate-rock aquifer was modeled as horizontally anisotropic, with the direction of maximum transmissivity aligned with the longitudinal axis of the valley. All lateral boundaries are no-flow boundaries. Recharge was applied uniformly to the topmost active layer with additional recharge added near the lateral boundaries to represent infiltration of runoff from adjacent crystalline-rock areas. The model was calibrated to June 2001 water levels in wells completed in the carbonate-rock aquifer, August 2000 stream base-flow measurements, and the approximate ground-water age.

The ground-water-flow model was constructed in part to test possible site contamination remediation alternatives. Four previously determined ground-water remediation alternatives (GW1, GW2, GW3, and GW4) were simulated. For GW1, the no-action alternative, simulated pathlines originating in the tetrachloroethene (PCE) and trichloroethene (TCE) source areas within the Ground-Water Contamination Site end at Pohatcong Creek near the confluence with Shabbecong Creek, although some particles went deeper in the aquifer system and ultimately discharge to Pohatcong Creek about 10 miles downvalley in Pohatcong Township. Remediation alternatives GW2, GW3, and GW4 include ground-water withdrawal, treatment, and reinjection. The design for GW2 includes wells in the TCE and PCE source areas that withdraw water at a total rate of 420 gallons per minutes (gal/min) and 100 gal/min, respectively. Flow-path analysis shows the system would capture all ground water within the 500-μg/L TCE isopleth and ground water from a small area in the PCE source area. The design for GW3 includes wells in the TCE and PCE source areas that withdraw and re-inject at a total rate of 1,400 gal/min and 420 gal/min, respectively. Flow-path analysis shows the system would capture all ground water within the 100-μg/L TCE isopleth in the TCE source area and all of the ground water in the estimated PCE source area. The design for GW4 includes

35 wells withdrawing a total of 10,820 gal/min. Most particles started in the flow-path analysis in the TCE source area and in an arbitrary area representing contamination farther downvalley were captured by the GW4 system, although a few particles traveled beneath the withdrawal wells and flowed down the valley, ultimately discharging to Pohatcong Creek.

Introduction

In 1978, the chlorinated solvents trichloroethene (TCE) and tetrachloroethene (PCE) were detected in Pohatcong Valley in production wells in Washington Borough and Washington Township, Warren County, New Jersey (CH2M Hill, 2005a). Subsequent investigation revealed that many domestic wells in Washington and Franklin Townships also were contaminated, and in 1989 the Pohatcong Valley Ground Water Contamination Site was added to the U.S. Environmental Protection Agency (USEPA) National Priority List (U.S. Environmental Protection Agency, 2006). A remedial investigation by the USEPA and CH2M Hill, with technical assistance from the U.S. Geological Survey (USGS), was begun in 1999. The USGS provided technical assistance to the USEPA on various hydrogeologic aspects of the remedial investigation. Results of the remedial investigation (including identification of a source area for each of the contaminants of concern, TCE and PCE) and feasibility of remedial alternatives are described in CH2M Hill (2005a) and CH2M Hill (2005b). USGS interpretation of aquifer test, stream base-flow, and water-level data are also presented in Carleton and others (2005). The simulation of ground-water flow in the Pohatcong Valley described here was conducted by the USGS, in cooperation with the USEPA.

The Pohatcong Valley, which is within the New England Province of New Jersey, is in an area of ridges (underlain by crystalline rocks) and valleys (underlain by carbonate rocks). The primary study area extends from Washington Borough and Washington Township to New Village in the southwestern part of Franklin Township (fig.1), but the model discussed in this report extends to the hydrologic boundaries and extends from near the headwaters of Pohatcong Creek to the Delaware River. Pohatcong Creek, a tributary to the Delaware River, drains a total area of 57 mi^2. The drainage area upstream from the USGS stream-gaging station Pohatcong Creek at New Village is 33 mi^2, and the mean annual discharge at New Village is about 32 ft^3/s. Land uses are industrial, commercial, and residential in settled areas such as Washington and Alpha Boroughs and the villages of Broadway, New Village, and Stewartsville, and agricultural and residential in the remaining areas of the valley.

This report describes the simulation of ground-water flow in the surficial and carbonate-rock aquifers of the Pohatcong Valley in Warren County, N.J., including all of the USEPA Pohatcong Valley Ground Water Contamination Site. The ground-water-flow model was used to estimate flow paths of ground-water from known sources of contamination in Washington Borough. The report describes aquifer and surface-water data and analyses required for model development. The report also includes a discussion of the hydrogeologic framework, aquifer hydraulic conductivity, stream base-flow estimates, ground-water withdrawals, ground-water age, and the flow-system water budget. The conceptual model and development of the finite-difference numerical model are described, inluding selection of the model grid and boundaries, the aquifer and boundary properties, and the calibration method. The resulting model parameters and flow paths for four ground-water remediation alternatives are presented.

Hydrogeology

The hydrogeology of the Pohatcong Valley is described in Carleton and others (2005) and CH2M Hill (2005a), and is summarized in the following sections. Additional material not included in previous reports is also included in the following sections.

Geology

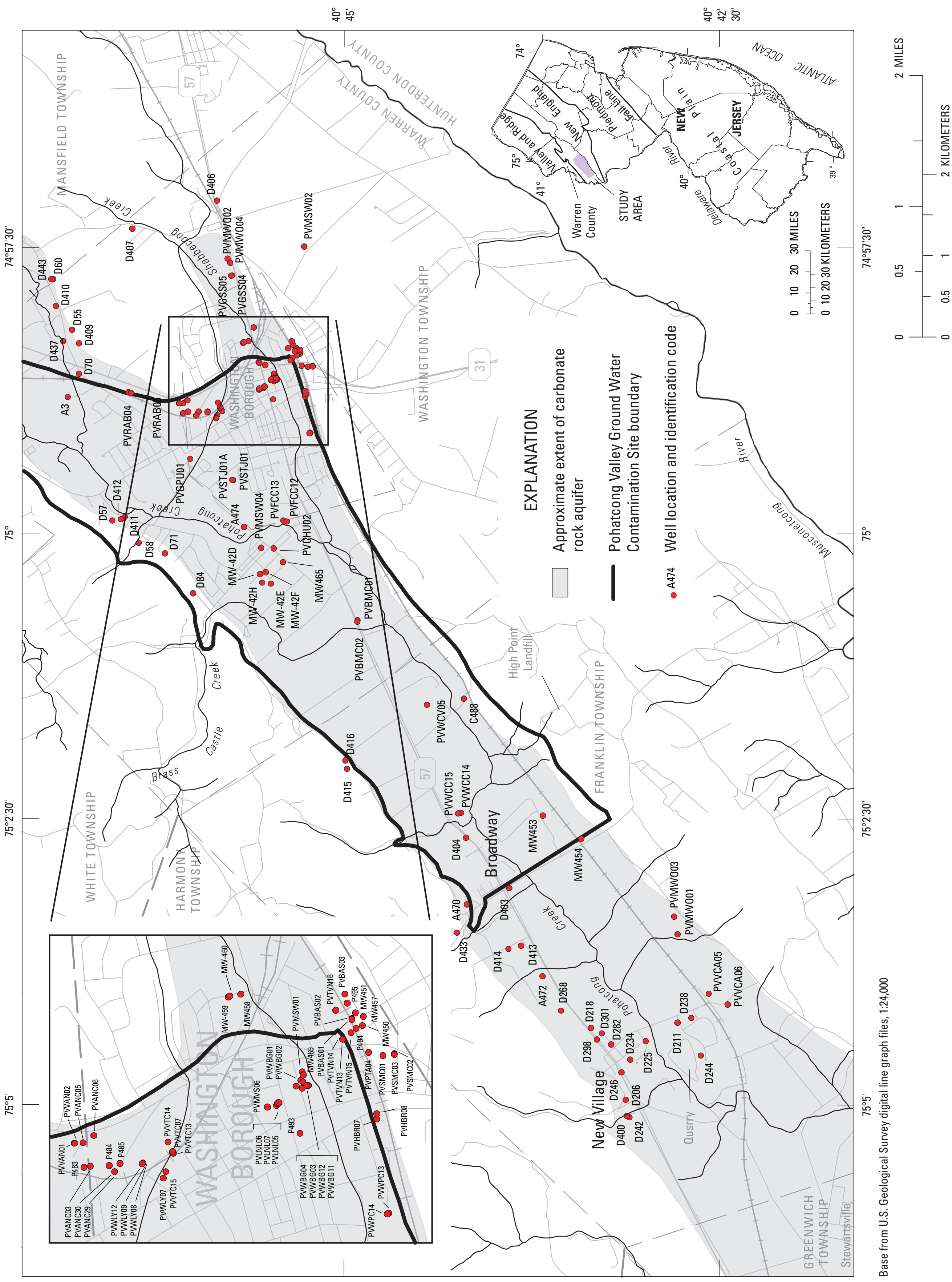
The Pohatcong Valley is bounded on the north and south sides by Middle Proterozoic (Precambrian) crystalline rocks. Although many rock types are included in this assemblage, for the purposes of this study they are considered to be one hydrogeologic unit.

The valley is underlain by Paleozoic rocks (Cambrian and Ordovician age, 600-435 million years old) including the Hardyston Quartzite (not shown in fig. 2), the Leithsville Formation, Allentown Dolomite, and Lower and Upper Parts of the Beekmantown Group, all members of the Kittatiny Supergroup, and Jacksonburg Limestone (fig. 2). All of the Paleozoic formations (except the Hardyston Quartzite) are carbonates, and these units are mapped and described by Drake and others (1994), Drake (1967), Drake and Lyttle (1985), and Drake and others (1996). Summaries of the geologic characteristics also can be found in Carleton and others (2005) and CH2M Hill (2005a).

The Leithsville Formation contains fine- to coarse-grained dolomite and calcitic dolomite, phyllite (between shale and mica schist), and thin beds of dolomite-cemented quartz. The unit is about 1,000 ft thick. Drake describes the Leithsville as one of the primary karst-forming formations in the study area (ICF-Kaiser, 1997, p. 8).

The Allentown Dolomite contains very fine- to medium-grained, rhythmically-bedded dolomite with beds and lenses of orthoquartzite, which are particularly abundant near the upper contact. Lower dolomite beds are interbedded with shaly dolomite. The shaly dolomite increases towards the conformable contact with the underlying Leithsville Formation. The Allentown Dolomite is about 1,900 ft thick.

The Beekmantown Group is divided into lower and upper parts. The Lower Part of the Beekmantown Group contains



Base from U.S. Geological Survey digital line graph files, 1:24,000

Figure 1. Location of selected wells in the Pohatcong Valley study area, Warren County, N.J.

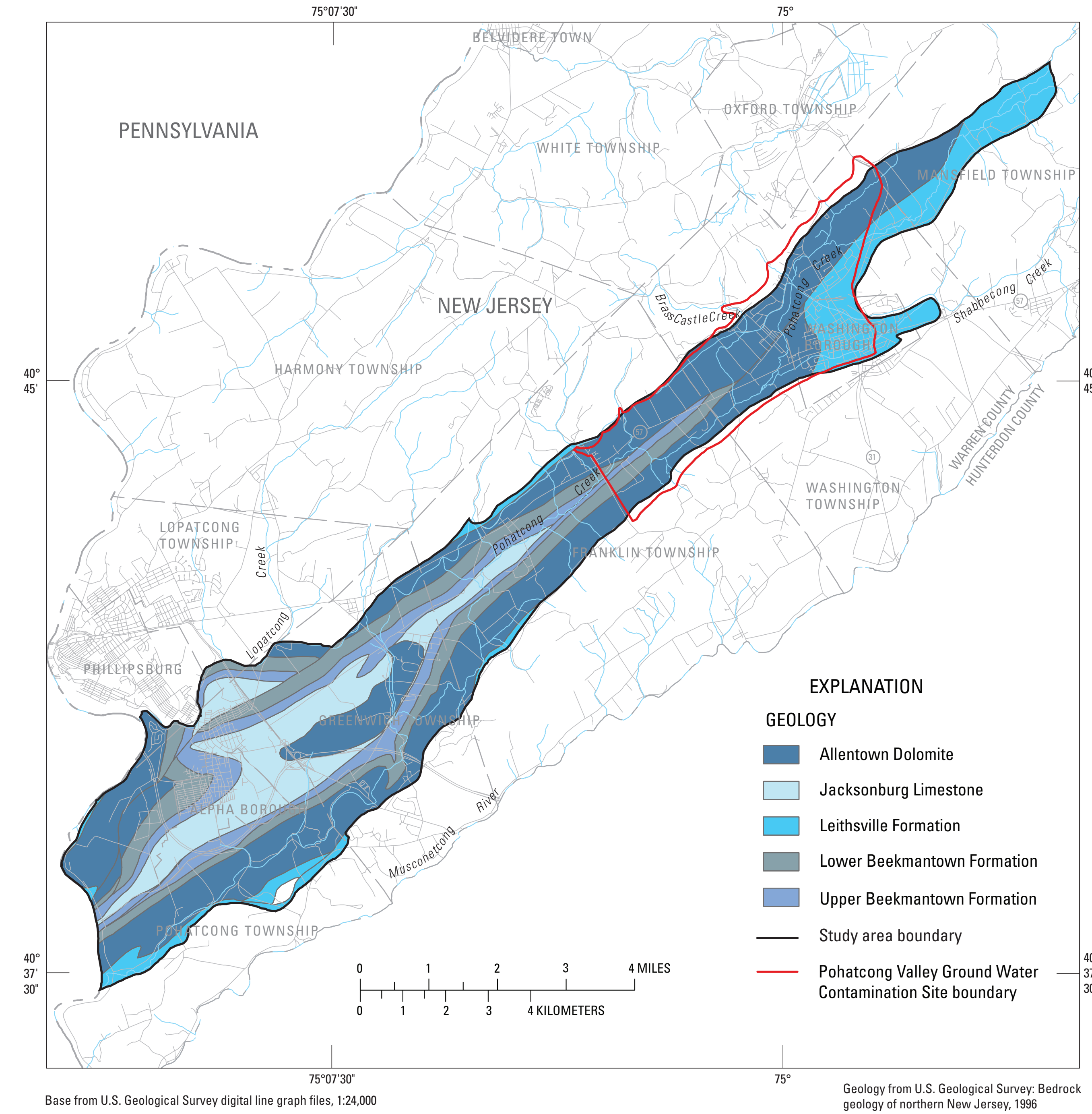


Figure 2. Carbonate-rock bedrock geology of the Pohatcong Valley study area, Warren County, N.J.

thin- to thick-bedded, very fine- to medium-grained dolomite, fine-grained limestone, dolomitic shale laminae surrounding limestone lenses, and solution-collapse breccia. The unit is as much as 700 ft thick. The Upper Part of the Beekmantown Group contains thin- to thick-bedded, very fine- to medium-grained dolomite. The upper beds of the unit locally contain medium-bedded, fine-grained limestone lenses. The lower beds contain chert lenses and locally occurring chert beds. The thickness of the upper part statewide ranges from near 0 to 800 ft; the unit could be as much as 500 ft thick in the Pohatcong Valley. Drake describes the Beekmantown Group as “the other [in addition to the Leithsville Formation] main karst unit in the study area” (ICF-Kaiser, 1997); however, sinkholes observed at land surface typically are present only in the Upper Part of the Beekmantown Group (Donald Monteverde, NJDEP, oral commun., 2001).

The Jacksonburg Limestone (fig. 2), present in the Pohatcong Valley only southwest of New Village (fig. 1), is divided into two units, the upper Cement Rock facies and lower Cement Limestone facies. The upper unit is a fine-grained, thin-bedded, argillaceous (clayey) limestone with some beds of crystalline limestone in places. The lower unit is a medium- to coarse-grained well-bedded calcarenite and fine- to medium-crystalline high-calcium limestone. The upper and lower units are from 300- to 600-ft thick and up to 200-ft thick, respectively.

The locations of contacts between the Paleozoic carbonate-rock formations are approximate in many locations within the Pohatcong Valley (fig. 2) because the rocks are covered by 30 ft or more of glacial, coalluvial, and alluvial deposits. The contacts between the carbonate (valley) and crystalline (ridge) rocks are regional thrust faults. The three faults mapped in the Pohatcong Valley are the Pohatcong (which joins the Kennedys fault just northeast of Washington Borough), Karrsville, and Brass Castle faults (fig. 3). The Pohatcong and Brass Castle faults form the southeastern and northwestern borders, respectively, of the carbonate rocks in the Pohatcong Valley. The locations of these faults are relatively well known because of the substantial change in rock type on either side of the fault, recognizable in drillers’ logs, outcrop, and float. The location of the Karrsville fault is less certain. The presence of the fault is indicated by otherwise unexplained sequences southwest of the study area (Donald Monteverde, NJDEP, oral commun., 2001) and low magnetic expression under Upper Pohatcong Mountain observed in aeromagnetic data (Drake and others, 1994).

The surficial deposits in the study area are composed primarily of deposits of Illinoian and Jerseyan age, recently re-worked alluvial sediments close to Pohatcong Creek, and weathered bedrock. The units mapped by Stanford and Ashley (1993) in the study area are mostly Jerseyan till plus one zone each of Jerseyan fluvial sediment, Illinoian moraine, and Illinoian till. Logs from wells throughout the study area typically describe low permeability, poorly sorted, clay-bearing sediments, but no clearly identifiable zone of high-permeability sediments.

Hydrology

The ground-water-flow system in the Pohatcong Valley area apparently has two distinct scales-- local recharge to, flow in, and discharge from the surficial aquifer to Pohatcong creek, and recharge to, and regional, downvalley flow in, the Cambrian/Ordovician carbonate formations that underlie the valley. Water also moves through the Precambrian crystalline rocks that bound the valley, but specific-capacity data indicate that the transmissivity and, therefore, the rate of flow is low compared to that in the carbonate rocks. Discharge from springs near the contact between the crystalline and carbonate rocks, and surface runoff from the hills, reenters the ground-water system as recharge to the surficial and carbonate-rock aquifers. Precipitation falling on the valley floor also becomes direct recharge to these aquifers (fig. 3).

The surficial unconsolidated sediments form a poorly productive, unconfined aquifer. No high-permeability zones have been documented and few domestic wells are completed in this aquifer. The thickness is highly variable, in part because of the substantial variations in the altitude of the top of the underlying competent carbonate rock over short horizontal distances (tens to hundreds of feet). A perched water table was observed in some locations; for example, in the center of Washington Borough, on the northern border of the borough, and near well PVCHU02 (table 1 at end of the report) west of the borough (fig. 1). In some locations, the bottom of the surficial deposits is unsaturated, including those where a perched water table is present, for example, on the northern border of Washington Borough. Data on water levels and depth to rock for various locations indicate that the water table is present within the surficial aquifer only where the bottom of the aquifer is below the regional water level in the underlying carbonate-rock aquifer and that horizontal flow in the surficial aquifer is a smaller component of the flow system than vertical transmission of recharge to the underlying carbonate-rock aquifer.

The unconfined to semi-confined carbonate-rock aquifer is made up of the Leithsville Formation, Allentown Dolomite, Beekmantown Group, and Jacksonburg Limestone. Specific-capacity data from records of wells completed in the Pohatcong Valley and from the USGS statewide Ground-Water Site Inventory (GWSI) database indicate that these formations have similar transmissivities and, therefore, are grouped together as one aquifer. The permeability of the aquifer is due primarily to secondary fractures, joints, and solution channels in the rock, and the aquifer is highly heterogeneous. Results of an aquifer test (Carleton and others, 2005) indicate that the aquifer is horizontally anisotropic: transmissivity along the axis of the valley is greater than transmissivity across the valley. The sources of this anisotropy most likely include fractures and solution channels preferentially created along bedding planes that strike along the axis of the valley (northeast/southwest) and dip steeply to the southeast. Fracture-orientation data from borehole acoustic televiewer logs of 16 boreholes are provided by CH2M Hill (2005a, app. D). Those logs show that most of

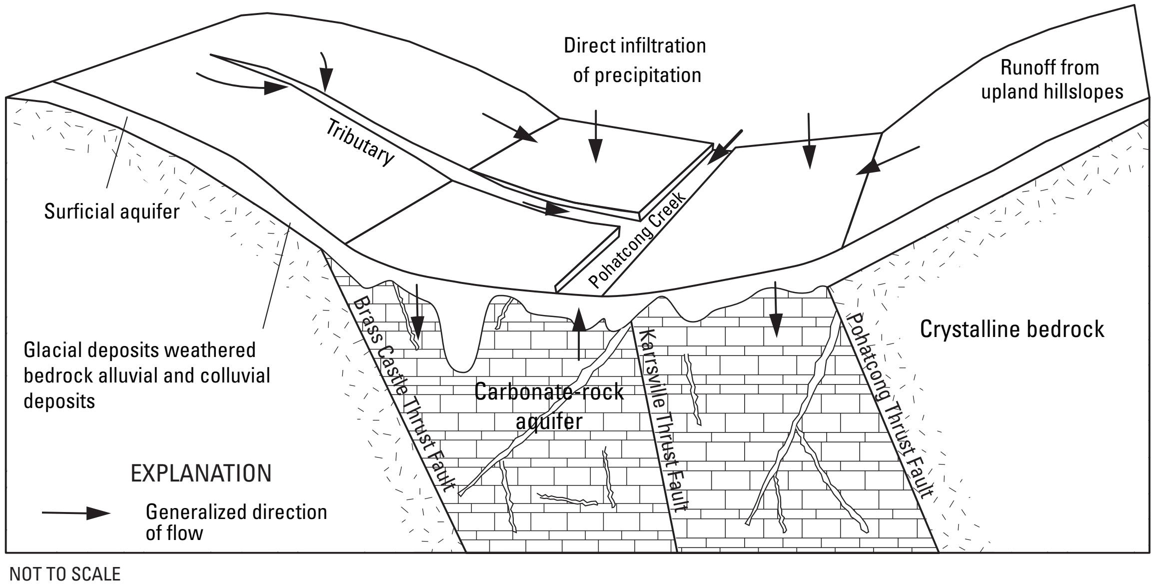


Figure 3. Cross section showing generalized geology and generalized flow directions, Pohatcong Valley, Warren County, N.J.

the fractures that dip at an angle greater than 30° have a strike in the range of 15° west of north to 105° east of north, indicating that bedding-plane fractures and fractures perpendicular to bedding but sharing the same strike are more prevalent than are other fracture sets. No major caves are known in the study area, but cavities up to three feet across were encountered during the drilling of wells PVLNL05 and PVWLY08, and a minor cave near the Delaware River has been reported (Richard Dalton, NJDEP, oral commun., 2006). Some collapse occurred in wells PVBMC01, PVGSS05, PVLNL05, PVRAB05, PVSTJ01, PVVAN01, PVWBG12, PVWCC15, and PVWLY08 during or after drilling. Well P483 (a re-injection well at an industrial facility) also is partly collapsed.

Depth to bedrock (which is equivalent to the thickness of the surficial aquifer) was estimated from geologists' logs for wells drilled as part of the Pohatcong Valley Ground Water Contamination Site study (CH2M Hill, 2005a), from other environmental investigations, and from drillers' logs for domestic wells (table 2). The carbonate formations commonly have a thick weathered zone, making identification of a precise bedrock/unconsolidated sediment interface difficult. For this report, the depth to bedrock noted in geologists' logs for monitoring wells drilled for this and other studies are considered reliable. Geologic information in drillers' logs for domestic wells may not be consistent, but the logs do provide an indication of depth where no other data are present. The depth to rock ranges from 20 feet in the central part of the valley to more than 240 feet in some areas and is highly variable over short distances. In some cases, the logs indicated both a depth to top of weathered rock and a depth to top of competent rock, so that the thickness of weathered rock could be estimated. The average and median thickness of weathered rock above each of the different carbonate formations was calculated to determine whether different formations had different weathering characteristics. The data indicate that the thickness of the weathered rock layer above the Leithsville Formation and Allentown Dolomite was less than that above the Beekmantown Group and Jacksonburg Limestone, but the subjective nature of the data do not support a strong conclusion.

The altitude of the top of bedrock (fig. 4) was calculated by subtracting the depth to rock from the land-surface altitude at each well. Land-surface altitudes are accurate to within 0.3 ft at most wells included in the water-level synoptic study (described below). The remaining land-surface altitudes were determined from a digital elevation model (DEM) with 30-meter horizontal resolution that, in turn, was derived from the 20-foot-interval contours on USGS topographic maps. Land-surface altitudes at 129 surveyed points were compared to the DEM altitudes for those points to estimate the accuracy of the DEM. The DEM altitudes differed by an average of 6.2 ft and as much as 20 ft from the surveyed altitudes. The average altitude of the top of bedrock increases from less than 300 ft near New Village to more than 500 ft in the northeastern end of the study area. The top of bedrock generally is lower in the central part of the valley than at the valley edges.

Fracture and Permeability Analysis

Ground-water flow in the carbonate formations is presumed to be in secondary openings of bedding-plane partings, joints, and faults, some of which have been enlarged by solution. On a site-specific scale, solution channels and sinkholes (karst features) may dominate the flow system. Drillers' logs commonly indicate that solution cavities/weathered zones in the carbonate-rock formations are filled with clay, which may reduce their importance to basin-wide transmissivity, but open

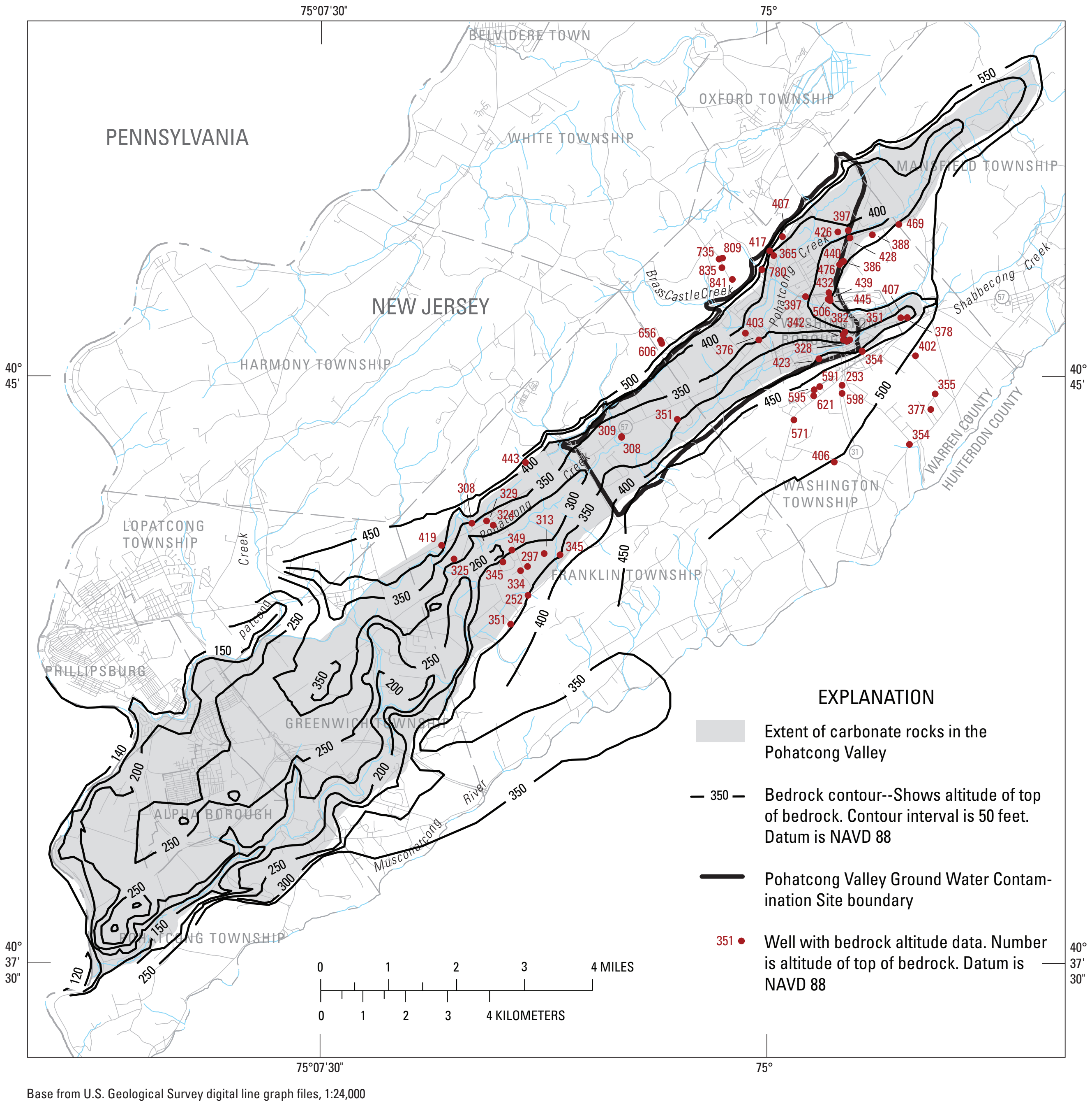


Figure 4. Estimated altitude of the top of bedrock, Pohatcong Valley, Warren County, N.J.

Table 2. Depth to top of weathered bedrock, estimated depth to top of competent bedrock, and thickness of weathered rock, Pohatcong Valley, Warren County, New Jersey

[Estimated thickness of weathered rock is depth to competent rock minus depth to top of rock; bls, below land surface]

Formation	Number of wells with data	Depth to top of weathered bedrock (in feet bls)		Depth to top of competent bedrock (in feet bls)		Thickness of weathered rock (in feet)	
		Mean	Median	Mean	Median	Mean	Median
Leithsville Formation	16	76	65	78	70	2	5
Allentown Dolomite	49	55	44	81	71	26	27
Beekmantown Group, Lower Part	28	49	50	109	81	60	31
Beekmantown Group, Upper Part	22	42	34	98	95	56	61
Jacksonburg Limestone	3	37	40	96	74	59	34

solution channels may locally increase the rate of ground-water flow. The orientation and extent of structural, solution-enlarged, bedding-plane, and other water-transmitting fractures are generally not known. The few measurements of rock outcrops in the area show strike along the northeast/southwest axis of the valley, (Drake and others, 1996), and dips to both southeast and northwest, usually at more than 45 degrees. The bedding strike implies that bedding-plane partings (and any associated higher permeability) are oriented along the axis of the valley.

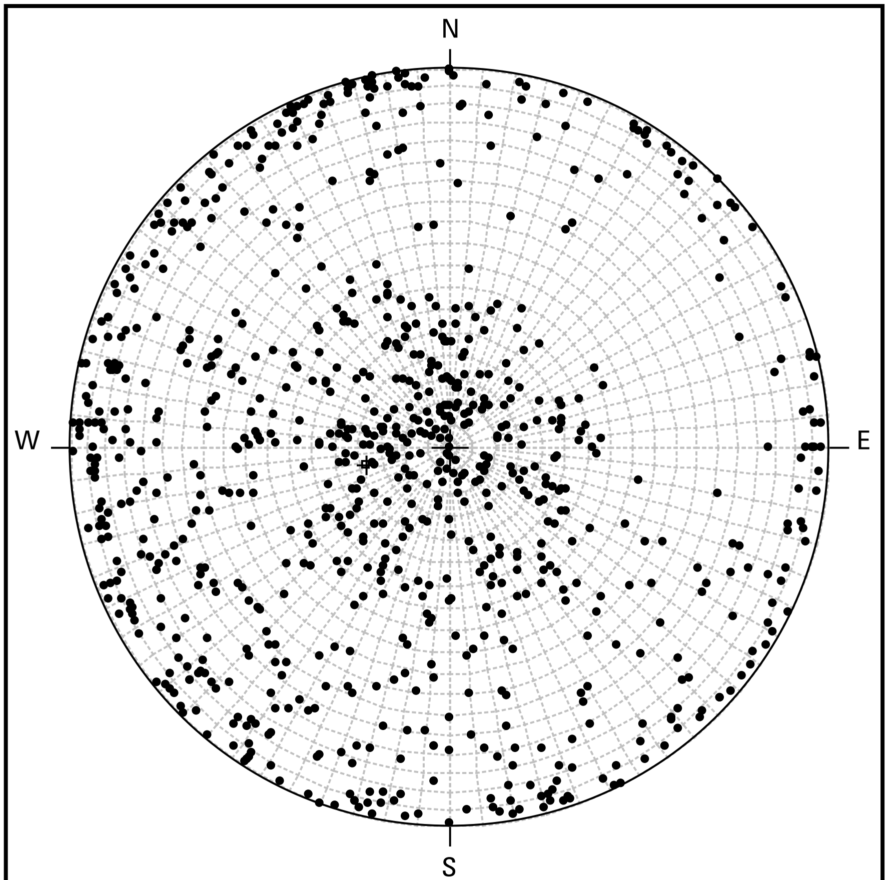
The permeability of the different geologic formations and any trends in fracture orientation were evaluated from aquifer-test data, specific-capacity data, and borehole-geophysical data. Estimates of transmissivity and hydraulic conductivity were calculated from data collected during aquifer tests conducted for this study (Carleton and others, 2005 and CH2M Hill, 2005a). Individual fractures in boreholes were detected with a borehole acoustic televiewer (which is magnetically oriented to north), and their orientations were estimated from the instrument output. The water-producing properties of the individual fractures is not known, but water-producing zones identified with the acoustic televiewer and heat-pulse flowmeter were isolated with inflatable packers and tested. Aquifer tests of zones isolated with inflatable packers led to transmissivity estimates for those individual zones, and the variation of aquifer properties with depth and location was quantified. Large-scale aquifer tests using the PVMSW01 and PVMSW04 production wells yielded data suitable for estimating the transmissivity of a larger volume of the aquifer in that area.

Analysis of Fracture Data from Borehole Acoustic Televiewer Logs

Thirteen boreholes open to carbonate rock were logged with an acoustic televiewer (CH2M-Hill, 2005a); little discernible pattern was observed to the strike/dip of fractures detected with the televiewer. The plunges (direction and magnitude of dip) of all 848 observed fractures in the 12 monitoring wells logged in 2000 (Phase I) and 2003 (Phase II) and in 1 domestic well logged for this study are shown in figure 5a, and strike directions are shown in figure 5b. Few fractures dip to the northeast, which is to be expected since the bedding planes dip to the southwest and southeast. Plunges are virtually uniformly distributed from angles of 0° to 90° and oriented in all directions other than northeast, indicating little bias to fracture orientation. A bias is apparent in the interpretation of Phase I versus Phase II data because far more sub-horizontal fractures were identified in Phase II than Phase I; the average dip of the Phase I fractures is 50° compared to 30° for Phase II. Determination of the direction of dip of fractures dipping less than about 30° is subject to more error than for the more steeply dipping fractures because the sinusoidal curve in the televiewer output is nearly horizontal, and therefore, the peak is difficult to pinpoint. The plunge for all fractures detected in the carbonate-rock wells with dips greater than 30 degrees are shown in figure 6a and strike directions are shown in figure 6b. Sixty percent of the fractures with dips greater than 30° strike northeast/southwest, indicating that bedding-plane fractures and fractures perpendicular to bedding planes but with the same or similar strike may be more prevalent.

Packer Test Analysis

Aquifer-isolation tests are conducted by separating a specific interval in a borehole with a packer and pumping that interval while monitoring water levels above, within, and below the pumped zone. Analysis of aquifer tests on isolated intervals in a well provides estimates of water-transmitting properties of specific zones. Tests conducted at multiple depths in multiple boreholes provide data useful for estimating the local heterogeneity of the aquifer and variations with depth and location. Analysis of results from multiple tests can provide estimates of regional aquifer properties. Aquifer tests were conducted in 47 isolated intervals of 15 boreholes



Lower Hemisphere Polar (Equal Area)
Data Plotted as Plunges
Total Number of Points = 848

⊕ Average azimuth and plunge are 258 and 71 degrees, respectively

Figure 5a. Stereoplot diagram for all fractures observed in thirteen wells drilled in carbonate formations, Pohatcong Valley, Warren County, N.J.

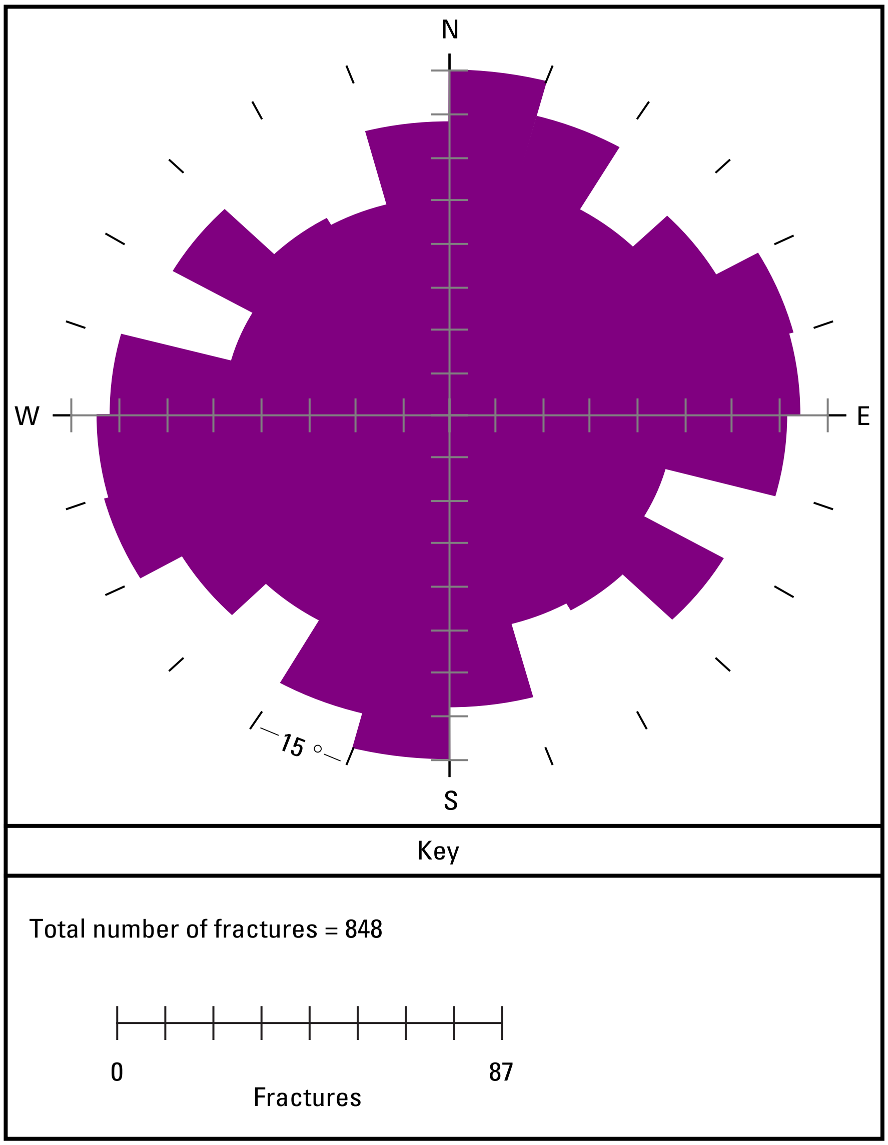


Figure 5b. Rose diagram showing strike for all fractures observed in thirteen wells drilled in carbonate formations, Pohatcong Valley, Warren County, N.J.

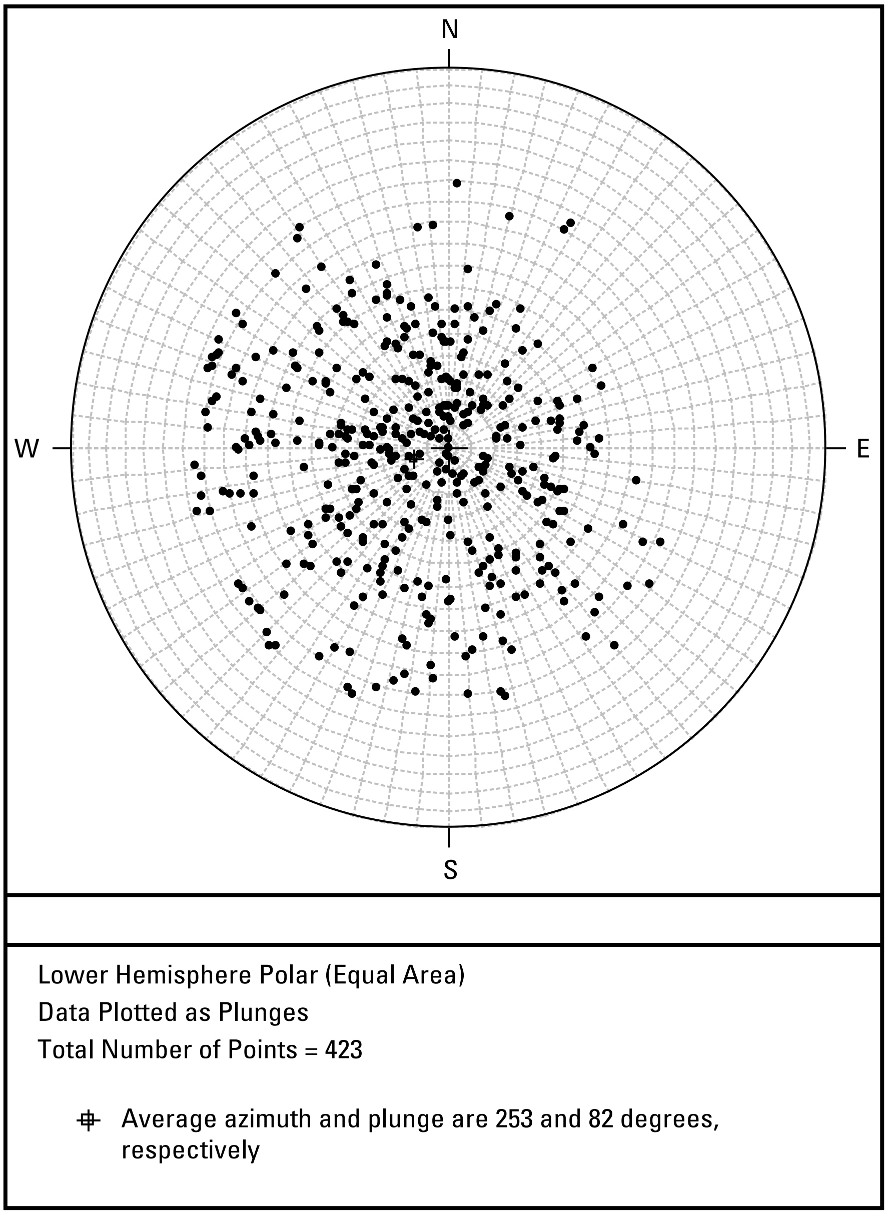


Figure 6a. Stereoplot diagram for all fractures with dips greater than or equal to 30 degrees observed in thirteen wells drilled in carbonate formations, Pohatcong Valley, Warren County, N.J.

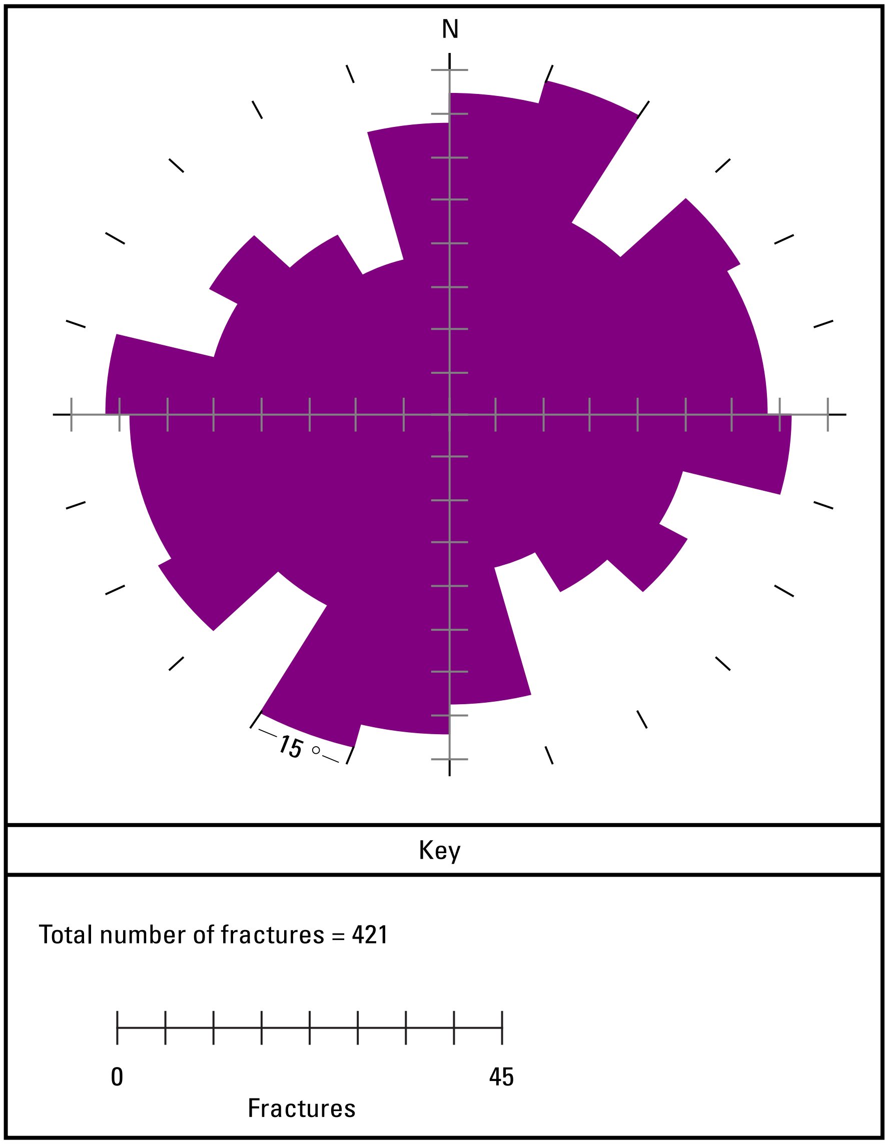


Figure 6b. Rose diagram showing strike for all fractures with dips greater than or equal to 30 degrees observed in thirteen wells drilled in carbonate formations, Pohatcong Valley, Warren County, N.J.

(CH2M-Hill, 2005a). For each borehole, geophysical logs (heat-pulse flowmeter, acoustic-televiewer, fluid-temperature and fluid-resistivity, and caliper) were analyzed to determine the location of water-producing zones. Two to five intervals were tested in each borehole. The intervals ranged in length from 7 to 113 feet; the median interval was 20 feet long. Each zone was pumped for 2 hours at the maximum sustainable rate up to about 50 gal/min unless the water-level drawdown was excessive, in which case the test was terminated when the water level was within about 5 feet of the pump. The water-level drawdown and recovery data were analyzed by several analytical methods, including the Theis (1935) non-leaky confined aquifer and recovery methods, the Hantush and Jacob (1955) leaky confined aquifer method, and the Cooper-Jacob (1946) straight-line confined-aquifer method. For each interval, the transmissivities estimated using the different analytical methods were averaged because the aquifer does not meet all of the simplifying assumptions of any method.

The highest estimated transmissivity, 37,000 ft^2/d, was calculated for the interval 240 to 260 feet below land surface (ft bls) in the St. Joseph's Church well (PVSTJ01A) (fig. 7). Estimated transmissivities were very high in one or more zones of several other wells, including PVWLY08, 18,713 ft^2/d at 180 to 210 ft bls; PVLNL05, 9,351 and 5,154 ft^2/d at 143 to 161 and 189 to 257 ft bls, respectively; and PVCHU02, 5,120 ft^2/d at 138 to 160 ft bls. Several intervals in boreholes drilled into the carbonate rock had estimated transmissivities of less than 10 ft^2/d, including PVWLY07, 2 ft^2/d at 176 to 196 ft bls, and PVBMC01, 2 ft^2/d at 34 to 60 ft bls and 4 ft^2/d at 60 to 80 ft bls. No trend of increasing or decreasing transmissivity with depth or higher or lower transmissivity in an area or particular formation was observed. The mean estimated transmissivity of all 47 intervals tested is 2,555 ft^2/d, and the median is 113 ft^2/d.

The intervals tested range in length from 7 to 113 feet; therefore, comparisons of hydraulic conductivities (transmissivity divided by length of interval tested) are useful. Horizontal hydraulic conductivities range from 0.05 ft/d to 1,845 ft/d, nearly five orders of magnitude. The mean is 137 ft/d and the median is 6 ft/d. The mean transmissivities and hydraulic conductivities are much higher than the median because of the few very high values. If the aquifer is assumed to have similar horizontal hydraulic conductivities over the assumed thickness of 500 ft, the mean transmissivity estimated from packer test hydraulic conductivities is 68,500 ft^2/d, and the median is 3,000 ft^2/d. However, it is unlikely that the highest permeabilities occur uniformly to depths of 500 ft.

Specific-Capacity Analysis

Specific capacity (SC), reported as gallons per minute per foot of drawdown ((gal/min)/ft), is an approximate measure that is proportional to the transmissivity of the aquifer. Specific-capacity data, although less useful than time-drawdown data, are commonly recorded by drillers on well-completion records and, therefore, are a valuable data source when other data are not available. Specific-capacity data for wells in the study area and for wells in the same formations throughout northern New Jersey for which data are stored in the USGS Ground Water Site Inventory (GWSI) database were analyzed to determine whether the horizontal hydraulic conductivities of the several carbonate formations are similar. The mean specific capacities of wells in the study area open to the Upper and Lower Parts of the Beekmantown Group and Allentown Dolomite are similar-- 2.7, 2.6, and 2.9 (gal/min)/ft with 10, 10, and 27 measurements, respectively (table 3a). The specific capacities confirm that these are transmissive formations. The medians are slightly lower than the means-- 2.0, 1.56, and 2.0 (gal/min)/ft, respectively. These means and medians are skewed to the low side of the true mean and median for all wells drilled in the study area because for many wells, drillers reported zero drawdown and, therefore, no specific capacity could be calculated. It is physically impossible to pump water from a well without causing some drawdown, however slight. Therefore, because the drawdowns were below the resolution of the driller's measuring technique, these wells clearly had a high specific capacity. The mean and median specific capacity values for 55 wells open to the crystalline rock is lower than those in the carbonate rocks-- 0.92 and 0.43, respectively. Two or fewer wells completed in the other carbonate-rock formations in the Pohatcong Valley have associated specific-capacity data.

A short borehole in a permeable formation can have the same specific capacity as a long borehole in a less permeable formation. Analyzing specific capacity per foot of opening makes it possible to directly compare the hydraulic conductivities of different formations. The median specific capacities per foot of open hole ([(gal/min)/ft]/ft) in the carbonate-rock Allentown Dolomite and Lower and Upper Parts of the Beekmantown Group are 0.11, 0.18, and 0.039 [(gal/min)/ft]/ft, respectively, and 0.0024 [(gal/min0/ft]/ft in the Precambrian crystalline rock. The one to two order-of-magnitude difference in specific capacity per foot of opening between the carbonate- and crystalline-rock aquifers shows the much lower permeability of the crystalline rocks. The lower permeability frequently causes drillers to complete wells in the crystalline rock with much longer open intervals than in carbonate rock to obtain suitable yields.

Mean specific capacities for wells throughout New Jersey completed in the Leithsville Formation, Allentown Dolomite, Epler Formation, Rickenbach Formation, and Kittatinny Limestone (all of the carbonate formations in the Paleozoic rocks) and listed in the USGS Ground-Water Site Inventory (GWSI) are 16, 19, 5.5, 0.76, and 13.6 (gal/min)/ft; the numbers of wells (n) are 80, 40, 4, 6, and 29, respectively. Median specific capacities are 1.7, 2.3, 5.3, 0.37, and 1.45 (gal/min)/ft, respectively. The medians are much lower because of the few outliers of very high-yielding wells.

Specific capacity ((gal/min)/ft) can be used to estimate transmissivity (typically reported in ft^2/day) or hydraulic conductivity (ft/day) (Heath, 1983, p. 60-61). Several assumptions were made, including that flow fits the assumptions of the

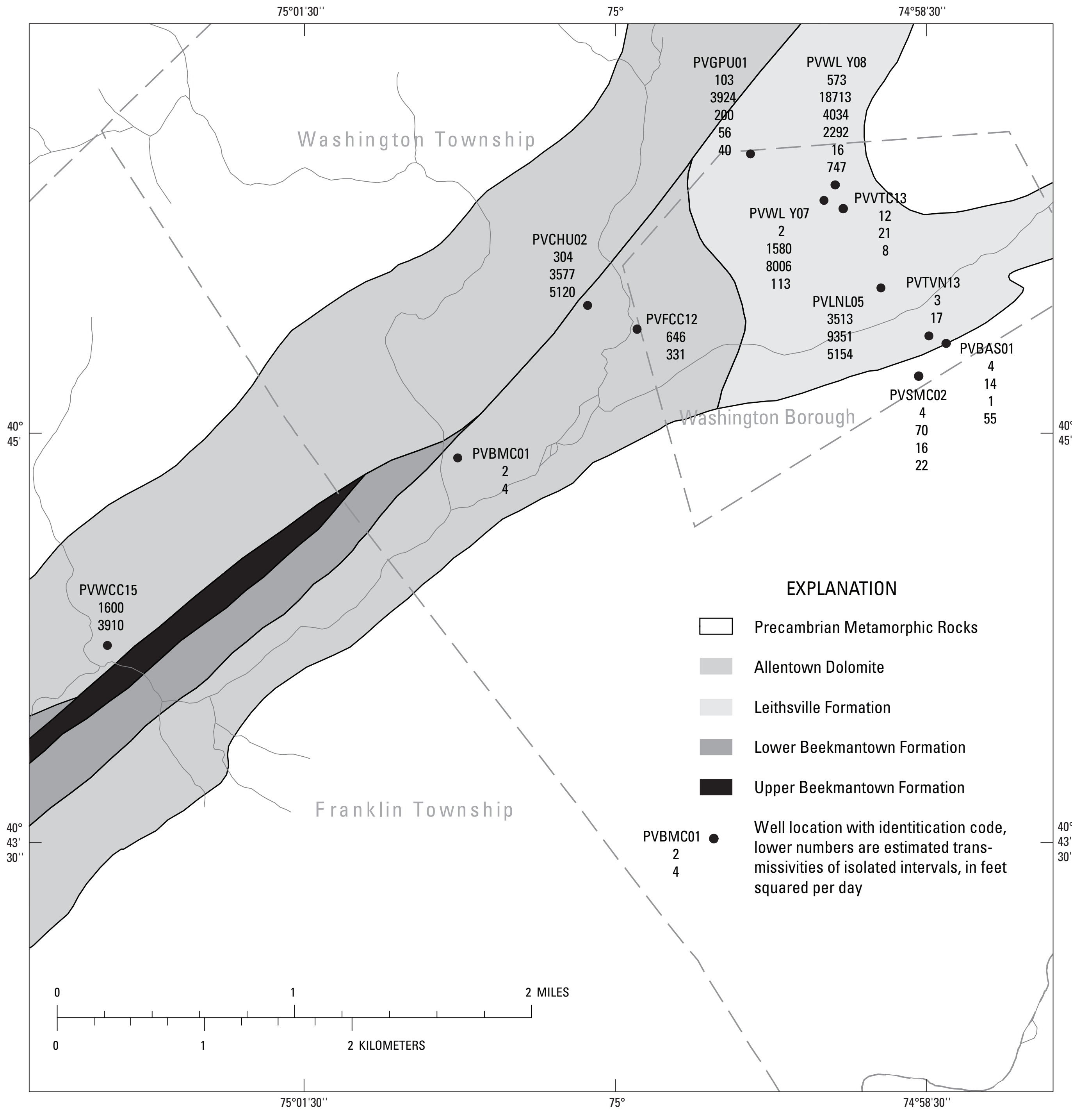


Figure 7. Location of selected wells for which transmissivities were estimated at specific intervals, Pohatcong Valley, Warren County, N.J.

Theis equation (including confined aquifer of infinite extent, isotropic flow, and no recharge or well loss), the open interval represents the full aquifer thickness, and the radius used in the Theis equation is the radius of the well. The Theis assumptions clearly were violated in this setting; however, the values provide an order-of-magnitude estimate of transmissivity and a useful starting point for model calibration. Transmissivity (defined as hydraulic conductivity times aquifer thickness) is converted to hydraulic conductivity by dividing by the open interval of the well. Hydraulic-conductivity estimates were calculated from specific-capacity data for 81 wells in the study area for which sufficient data were available (table 3a). For study-area wells completed in the Allentown Dolomite, in the Upper and Lower Parts of the Beekmantown Group, and in Precambrian rocks, median hydraulic conductivities estimated from specific-capacity data are 11, 43, 47, and 0.50 ft/day, respectively. As mentioned above, these medians are biased low because, for many wells, zero drawdown was reported and specific capacity could not be calculated.

No slug tests or other hydraulic tests were conducted in the surficial aquifer. However, well-sampling logs (Brian Wied, CH2M Hill, written commun., 2004) for surficial-aquifer wells that included pumping rate and water-level drawdown were used to calculate specific capacity. The specific capacity data were analyzed, as described above, to estimate transmissivity and hydraulic conductivity (table 3b). The median hydraulic conductivity is 2.7 ft/d.

Aquifer-Test Analysis

Carleton and others (2005) conducted several large-scale aquifer tests (24 hours or longer) by pumping production wells PVMSW01 and PVMSW04 and measuring the aquifer response in nearby observation wells. Results from the aquifer tests of well PVMSW01, completed in the Leithsville Formation and open from 88 to 345 ft below land surface, indicate the transmissivity, hydraulic conductivity, and storage coefficient of the tested part of the aquifer are about 8,600 ft^2/d, 34 ft/d, and 1×10^{-2}, respectively (Carleton and others, 2005). Transmissivities calculated from the aquifer tests of well PVMSW04, completed in the Allentown Dolomite and open from 143 to 184 ft below land surface, range from 2,500 to 9,600 ft^2/d (horizontal hydraulic conductivity ranges from 61 ft/d to 234 ft/d), and calculated storage coefficients range from 7×10^{-3} to 3×10^{-2}. Analysis of the horizontal anisotropy of the aquifer in the vicinity of PVMSW04 indicated the approximate maximum and minimum transmissivities are 7,600 and 3,500 ft^2/d (hydraulic conductivity 190 ft/d and 85 ft/d), respectively. The direction of maximum transmissivity is N.58°E., and the ratio of anisotropy (maximum transmissivity: minimum transmissivity) is 2.2:1. The direction of maximum transmissivity, N. 58°E., is similar to the trend of the valley (about N. 55°E.). Most strike measurements in Pohatcong Valley reported by Drake and others (1996) are close to the axis of the valley, indicating that bedding-plane fractures and structural fractures that strike parallel to bedding planes could provide a preferential flow path (Carleton and others, 2005).

Stream Base Flow

Measurement of streamflow under low-water (base-flow) conditions provides an estimate of ground-water discharge from the aquifer system to streams. Because ground-water recharge is difficult to measure, it can be equated to basin-wide stream base flow additions (such as sewage-treatment outflows) plus withdrawals or other discharges (such as consumptive-use ground-water withdrawals) from the system. Ground-water flow into or out of a basin also must be subtracted or added, respectively, when estimating recharge from stream base flow.

Synoptic measurements of stream base flow were made by Carleton and others (2005) on Pohatcong Creek, Shabbecong Creek, Brass Castle Creek, and nine smaller tributaries on July 13 and August 8, 2000 (fig. 8). Data from both base-flow surveys indicate that Pohatcong Creek is a steadily gaining stream along the measured reaches, with much of the gain coming from tributaries that originate in the area underlain by Precambrian crystalline rocks. Brass Castle Creek lost flow to the underlying formations over the relatively long reach (compared to other streams in the study area) from the crystalline/carbonate rock contact to the confluence with Pohatcong Creek. Shabbecong Creek either was losing flow (July 13 measurement) or had neither gain nor loss (August 8 measurement) from the crystalline/carbonate rock contact to the first measurement point on the reach within the area underlain by carbonate rock. Downstream, it was a gaining stream. Some or all of the downstream gain was effluent from the sewage-treatment plant in Washington Borough. Measurements at an upstream station and a downstream station on three tributaries to Pohatcong Creek show neither a clear gaining or losing pattern: tributary 9 was gaining flow between measurement points on both July 13 and August 8, tributary 10 was losing flow on July 13 but showed no significant change in flow on August 8, and tributary 7 showed no significant change in flow on August 8 (no measurement was made at one of the stations on tributary 7 on July 13). Flows in several small tributaries to Pohatcong Creek are affected by diversions into small agricultural ponds in their upstream reaches. The ponds typically lose water to the surficial aquifer.

Pohatcong Creek is underlain by carbonate rocks with karst features; therefore, it was considered possible that one or more springs account for a large percentage of ground-water discharge to the creek. The base-flow measurements made in July and August 2000 show that Pohatcong Creek steadily gained flow, but most of the gain downstream from station 01455145 (just above the confluence with Shabbecong Creek) was from tributaries in which virtually all flow originated in the crystalline rock areas (valley walls). Therefore, it was concluded that no springs contributed large flows to Pohatcong Creek under the streamflow and ground-water level conditions of July 13 or August 8, 2000. Surface- and ground-water levels

Table 3a. Specific capacity, transmissivity, and hydraulic conductivity data for selected bedrock wells, Pohatcong Valley, Warren County, New Jersey

[ft bls, feet below land surface; gal/min, gallons per minute; (gal/min)/ft, gallons per minute per foot of drawdown; hrs, hours; in., inches; ft/d, feet per day; ft^2/d, square feet per day; u, Theis variable; W(u), Theis well function; Ch, Hardyston Quartzite; Cl, Leithsville Formation; Obl, Lower Beekmantown Formation; Obu, Upper Beekmantown Formation; OCa, Allentown Dolomite; PreC, PreCambrian]

Map identification code	Geologic formation	Depth to top of open interval (ft bls)	Depth to bottom of open interval (ft bls)	Specific capacity ((gal/min)/ft)	Duration of pumping (hrs)	Well diameter (in.)	Estimated hydraulic conductivity for calculation (ft/d)	Estimated transmissivity for calculation (ft^2/d)	1/u	W(u)	Calculated transmissivity (ft^2/d)	Calculated hydraulic conductivity (ft/d)
D265	Ch	175	224	2.5	4	6	12.56	615.44	3.28E+07	16.74	641.46	13.09
D55	Ch	102	124	0.8	4	6	12.56	276.32	1.47E+07	15.9	194.97	8.86
D208	Ch	101	248	0.533	4	6	12.56	1846.32	9.85E+07	17.84	145.84	0.99
D93	Ch	136	295	0.212	4	6	12.56	1997.04	1.07E+08	17.95	58.36	0.37
I1A	Cl	140	260	11.029	8	8	12.94	1552.8	9.32E+07	17.84	3015.91	25.13
D294	Obl	62	65	5.714	4	6	12.94	38.82	2.07E+06	13.93	1220.07	406.69
D212	Obl	163	166	2.857	4	6	12.94	38.82	2.07E+06	13.93	610.03	203.34
D230	Obl	244	247	1.471	4	6	12.94	38.82	2.07E+06	13.93	313.99	104.66
D220	Obl	225	228	1.333	4	6	12.94	38.82	2.07E+06	13.93	284.68	94.89
D209	Obl	222	228	1.667	4	6	12.94	77.64	4.14E+06	14.62	373.48	62.25
D301	Obl	168	225	5	4	6	12.94	737.58	3.93E+07	16.93	1297.47	22.76
D299	Obl	60	142	4	4	6	12.94	1061.08	5.66E+07	17.31	1061.28	12.94
D251	Obl	0	80	2.353	4	6	12.94	1035.2	5.52E+07	17.31	624.28	7.80
D282	Obl	74	103	0.667	4	6	12.94	375.26	2.00E+07	16.23	165.84	5.72
D223	Obl	110	215	1.5	4	6	12.94	1358.7	7.25E+07	17.58	404.19	3.85
D211	Obu	121	123	1.5	4	6	12.94	25.88	1.38E+06	13.6	312.68	156.34
D252	Obu	63	75	2.5	4	6	12.94	155.28	8.28E+06	15.28	585.51	48.79
D219	Obu	132	140	1.667	4	6	12.94	103.52	5.52E+06	15.01	383.44	47.93
D205	Obu	98	108	2	4	6	12.94	129.4	6.90E+06	15.28	468.41	46.84
D213	Obu	167	185	0.5	4	6	12.94	232.92	1.24E+07	15.76	120.78	6.71

Table 3a. Specific capacity, transmissivity, and hydraulic conductivity data for selected bedrock wells, Pohatcong Valley, Warren County, New Jersey—Continued

[ft bls, feet below land surface; gal/min, gallons per minute; (gal/min)/ft, gallons per minute per foot of drawdown; hrs, hours; in., inches; ft/d, feet per day; ft^2/d, square feet per day; u, Theis variable; W(u), Theis well function; Ch, Hardyston Quartzite; Cl, Leithsville Formation; Obl, Lower Beekmantown Formation; Obu, Upper Beekmantown Formation; OCa, Allentown Dolomite; PreC, PreCambrian]

Map identification code	Geologic formation	Depth to top of open interval (ft bls)	Depth to bottom of open interval (ft bls)	Specific capacity ((gal/min)/ft)	Duration of pumping (hrs)	Well diameter (in.)	Estimated hydraulic conductivity for calculation (ft/d)	Estimated transmissivity for calculation (ft^2/d)	1/u	W(u)	Calculated transmissivity (ft^2/d)	Calculated hydraulic conductivity (ft/d)
D210	Obu	125	208	2	4	6	12.94	1074.02	5.73E+07	17.31	530.64	6.39
D82	Obu	51	225	0.214	4	6	12.94	2251.56	1.20E+08	18.07	59.35	0.34
D253	OCa	129	137	10	4	6	12.94	103.52	5.52E+06	15.01	2300.66	287.58
D206	OCa	170	172	2.273	4	6	12.94	25.88	1.38E+06	13.6	473.76	236.88
D295	OCa	151	164	3.333	4	6	12.94	168.22	8.97E+06	15.54	793.96	61.07
D261	OCa	138	150	2	4	6	12.94	155.28	8.28E+06	15.28	468.41	39.03
D302	OCa	192	208	1.538	4	6	12.94	207.04	1.10E+07	15.65	369.04	23.06
D207	OCa	106	128	2	4	6	12.94	284.68	1.52E+07	15.9	487.41	22.16
D96	OCa	127	165	3.333	4	6	12.94	491.72	2.62E+07	16.46	840.97	22.13
D81	OCa	93	145	3.182	4	6	12.94	672.88	3.59E+07	16.74	816.40	15.70
D298	OCa	115	166	3.077	4	6	12.94	659.94	3.52E+07	16.74	789.48	15.48
D305	OCa	81	100	1.111	4	6	12.94	245.86	1.31E+07	15.76	268.40	14.13
D85	OCa	86	145	2.857	4	6	12.94	763.46	4.07E+07	16.93	741.41	12.57
D18	OCa	58	104	2.143	4	6	12.94	595.24	3.17E+07	16.74	549.82	11.95
A101B	OCa	102	300	7.692	4	6	12.94	2562.12	1.37E+08	18.2	2145.85	10.84
D254	OCa	96	303	8	4	6	12.94	2678.58	1.43E+08	18.2	2231.68	10.78
A101C	OCa	91	300	7.333	4	6	12.94	2704.46	1.44E+08	18.2	2045.71	9.79
D79	OCa	61	100	1.25	8	6	12.94	504.66	5.38E+07	17.15	328.58	8.43
D284	OCa	92	122	1	4	6	12.94	388.2	2.07E+07	16.23	248.77	8.29
D14	OCa	165	180	0.388	4	6	12.94	194.1	1.04E+07	15.54	92.32	6.15

Table 3a. Specific capacity, transmissivity, and hydraulic conductivity data for selected bedrock wells, Pohatcong Valley, Warren County, New Jersey—Continued

[ft bls, feet below land surface; gal/min, gallons per minute; (gal/min)/ft, gallons per minute per foot of drawdown; hrs, hours; in., inches; ft/d, feet per day; ft^2/d, square feet per day; u, Theis variable; W(u), Theis well function; Ch, Hardyston Quartzite; Cl, Leithsville Formation; Obl, Lower Beekmantown Formation; Obu, Upper Beekmantown Formation; OCa, Allentown Dolomite; PreC, PreCambrian]

Map identification code	Geologic formation	Depth to top of open interval (ft bls)	Depth to bottom of open interval (ft bls)	Specific capacity ((gal/min)/ft)	Duration of pumping (hrs)	Well diameter (in.)	Estimated hydraulic conductivity for calculation (ft/d)	Estimated transmissivity for calculation (ft^2/d)	1/u	W(u)	Calculated transmissivity (ft^2/d)	Calculated hydraulic conductivity (ft/d)
D300	OCa	74	145	1.364	4	6	12.94	918.74	4.90E+07	17.15	358.45	5.05
D57	OCa	131	248	1.538	4	6	12.94	1513.98	8.07E+07	17.58	414.55	3.54
A3	OCa	67	200	0.4	1	6	12.94	1721.02	2.29E+07	16.34	100.18	0.75
D293	OCa	60	290	0.455	3	6	12.94	2976.2	1.19E+08	18.07	125.89	0.55
D119	OCa	54	175	0.167	4	6	12.94	1565.74	8.35E+07	17.58	44.91	0.37
D13	OCa	53	175	0.154	4	6	12.94	1578.68	8.42E+07	17.58	41.45	0.34
I8A	OCa	103	599	0.358	7	8	12.94	6418.24	3.37E+08	19.05	104.41	0.21
I2A	OCa	74	496	0.252	8	8	12.94	5460.68	3.28E+08	19.05	73.56	0.17
D260	Oj	84	100	1.5	4	6	12.94	207.04	1.10E+07	15.65	359.81	22.49
D40	PreC	50	108	8	4	6	0.5	29	1.55E+06	13.6	1667.63	28.75
D11	PreC	50	128	5	4	6	0.5	39	2.08E+06	13.93	1067.56	13.69
D116	PreC	58	125	2	4	6	0.5	33.5	1.79E+06	13.75	421.51	6.29
D92	PreC	122	188	0.667	4	6	0.5	33	1.76E+06	13.75	140.50	2.13
D12	PreC	71	295	0.033	4	6	0.5	112	5.97E+06	15.01	7.67	0.03
D90	PreC	63	408	0.02	4	6	0.5	172.5	9.20E+06	15.54	4.67	0.014
D201	PreC	50	180	3.333	1	6	0.5	65	8.67E+05	12.98	663.17	5.10
D83	PreC	50	228	0.444	4	6	0.5	89	4.75E+06	14.85	101.16	0.57
D89	PreC	50	248	0.48	4	6	0.5	99	5.28E+06	14.85	109.25	0.55
D54	PreC	50	300	0.5	2	6	0.5	125	3.33E+06	14.44	110.66	0.44
D264	PreC	104	400	0.24	4	6	0.5	148	7.89E+06	15.28	56.21	0.19

Table 3a. Specific capacity, transmissivity, and hydraulic conductivity data for selected bedrock wells, Pohatcong Valley, Warren County, New Jersey—Continued

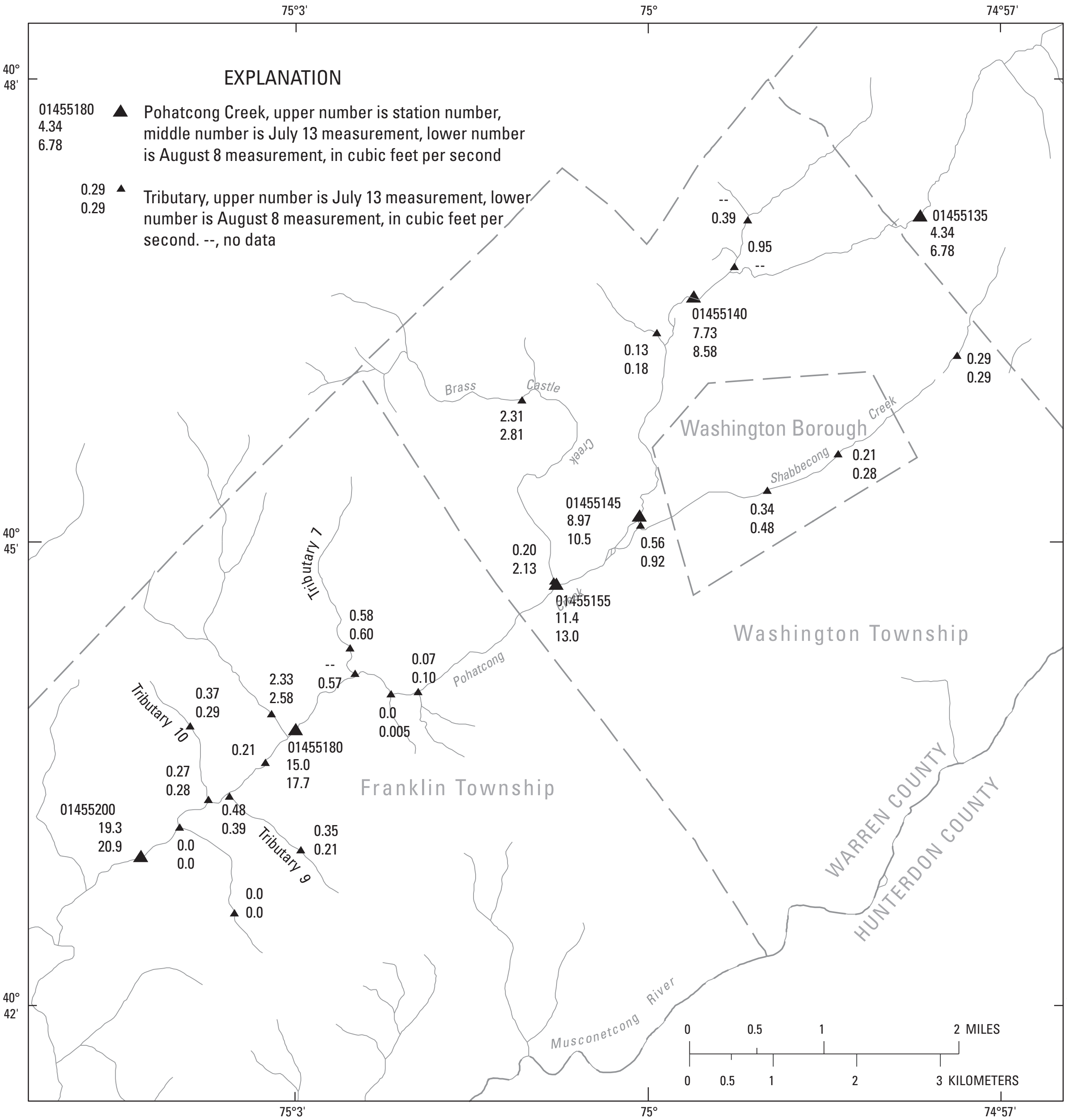
[ft bls, feet below land surface; gal/min, gallons per minute; (gal/min)/ft, gallons per minute per foot of drawdown; hrs, hours; in., inches; ft/d, feet per day; ft^2/d, square feet per day; u, Theis variable; W(u), Theis well function; Ch, Hardyston Quartzite; Cl, Leithsville Formation; Obl, Lower Beekmantown Formation; Obu, Upper Beekmantown Formation; OCa, Allentown Dolomite; PreC, PreCambrian]

Map identification code	Geologic formation	Depth to top of open interval (ft bls)	Depth to bottom of open interval (ft bls)	Specific capacity ((gal/min)/ft)	Duration of pumping (hrs)	Well diameter (in.)	Estimated hydraulic conductivity for calculation (ft/d)	Estimated transmissivity for calculation (ft^2/d)	1/u	W(u)	Calculated transmissivity (ft^2/d)	Calculated hydraulic conductivity (ft/d)
D163	PreC	50	225	0.118	5	6	0.5	87.5	5.83E+06	15.01	27.07	0.15
D148	PreC	52	125	0.333	5	6	0.5	36.5	2.43E+06	14.15	72.29	0.99
D152	PreC	63	170	0.058	5	6	0.5	53.5	3.57E+06	14.44	12.91	0.12
D6	PreC	62	180	0.769	1	6	0.5	59	7.87E+05	12.98	153.04	1.30
D91	PreC	98	205	0.35	4	6	0.5	53.5	2.85E+06	14.29	76.66	0.72
D4	PreC	52	320	0.8	1	6	0.5	134	1.79E+06	13.75	168.60	0.63
D1	PreC	50	260	0.5	1	6	0.5	105	1.40E+06	13.6	104.23	0.50
D2	PreC	56	360	0.031	1	6	0.5	152	2.03E+06	13.93	6.67	0.02
D8	PreC	70	300	0.019	1	6	0.5	115	1.53E+06	13.6	3.97	0.017
D303	PreC	97	125	2	4	6	0.5	14	7.47E+05	12.98	397.90	14.21
D262	PreC	79	275	0.368	4	6	0.5	98	5.23E+06	14.85	83.86	0.43
D263	PreC	49	300	0.429	1	6	0.5	125.5	1.67E+06	13.75	90.32	0.36
D16	PreC	128	303	0.182	4	6	0.5	87.5	4.67E+06	14.85	41.38	0.24
D26	PreC	110	188	1	4	6	0.5	39	2.08E+06	13.93	213.51	2.74
D87	PreC	87	200	0.8	8	6	0.5	56.5	6.03E+06	15.01	184.05	1.63
D257	PreC	49	168	0.6	4	6	0.5	59.5	3.17E+06	14.44	132.80	1.12
D27	PreC	90	295	0.5	4	6	0.5	102.5	5.47E+06	15.01	115.03	0.56
D53	PreC	129	340	0.4	4	6	0.5	105.5	5.63E+06	15.01	92.03	0.44
D255	PreC	72	268	0.333	4	6	0.5	98	5.23E+06	14.85	75.87	0.39
D307	PreC	147	333	0.143	4	6	0.5	93	4.96E+06	14.85	32.52	0.17
D258	PreC	79	460	0.009	4	6	0.5	190.5	1.02E+07	15.54	2.13	0.006

Table 3b. Specific capacity, transmissivity, and hydraulic conductivity data for selected surficial-aquifer wells, Pohatcong Valley, Warren County, New Jersey

[ft, feet; (gal/min) /ft, gallons per minute per foot of drawdown; ft/d, feet per day; ft^2/d, feet squared per day; u, Theis variable; W(u), Theis well function; storage coefficient estimated to be 0.01 for calculation of 1/u]

Map identification code	Screen length (ft)	Specific capacity ((gal/min)/ft)	Duration of test (minutes)	Borehole diameter (inches)	Estimated hydraulic conductivity for calculation (ft/d)	Estimated transmissivity for calculation (ft^2/d)	1/u	W(u)	Calculated transmissivity (ft^2/d)	Calculated hydraulic conductivity (ft/d)
PVANC39	10	0.065	30	6	0.60	6	8.03E+02	6.07	6	0.6
PVBAS02	10	1.1	25	6	15.42	154	1.71E+04	9.14	154	15
PVBAS03	10	0.78	25	6	10.41	104	1.16E+04	8.74	104	10
PVBMC02	10	1	30	6	14.53	145	1.94E+04	9.33	143	14
PVFCC13	10	2.6	50	6	44.30	443	9.84E+04	10.94	436	44
PVGSS04	10	0.044	35	6	0.39	4	6.09E+02	5.8	4	0.39
PVHBR07	13	0.25	25	6	2.26	29	3.26E+03	7.53	29	2.3
PVHBR08	13	0.096	70	6	0.85	11	3.44E+03	7.53	11	0.85
PVLNL06	10	2.1	45	6	33.67	337	6.73E+04	10.41	335	33
PVLNL07	10	0.17	5	6	1.35	13	3.00E+02	5.08	13	1.3
PVMVS06	10	2.2	20	6	32.67	327	2.90E+04	9.68	326	32
PVVAN02	10	0.089	29	6	0.88	9	1.13E+03	6.44	9	0.88
PVVTC15	10	0.21	17	6	2.20	22	1.66E+03	6.84	22	2.2
PVWBG01	10	0.065	21	8	0.51	5	2.66E+02	4.95	5	0.51
PVWBG03	13	0.024	43	8	0.13	2	1.86E+02	4.73	2	0.13
PVWBG11	10	0.033	35	6	0.28	3	4.29E+02	5.42	3	0.28
PVWCC14	10	5.3	65	6	97.11	971	2.81E+05	11.99	974	97
PVWLY09	10	0.29	15	6	3.09	31	2.06E+03	7.02	31	3.1
PVWPC14	20	1.5	35	6	11.38	228	3.54E+04	9.84	226	11
PVRAB04	10	2.6	60	6	45.19	452	1.21E+05	11.16	445	45



Base from U.S. Geological Survey digital line graph files, 1:24,000

Figure 8. Location of stream base-flow measurement stations on Pohatcong Creek and selected tributaries, Warren County, N.J., July 13 and August 8, 2000.

measured during the period of low ground-water levels in June 2002 (Carleton and others, 2005) indicate that, at all measured locations, surface-water levels were higher than ground-water levels in the underlying carbonate-rock aquifer. Therefore, it is likely that during periods of low ground-water levels, Pohatcong Creek does not gain flow from the carbonate-rock aquifer and probably loses flow to the surficial aquifer (which subsequently recharges the carbonate-rock aquifer).

Stream-Hydrograph Separation

Streamflow consists of two components: direct runoff and base flow. Direct runoff diminishes to zero within a few days after a precipitation event. Base flow is ground-water discharge to the stream and, in the absence of recharge to ground water from precipitation, diminishes slowly as water is removed from aquifer storage. The base-flow-separation technique PART described by Rutledge (1993) partitions stream discharge into direct-runoff and base-flow components using stream-discharge data from continuous-record gaging stations as input. Base-flow separation analyses were done for four continuous-record gaging stations located outside the Pohatcong drainage basin, on Pequest River, Musconetcong River, Paulins Kill, and Flat Brook, and for two continuous gaging stations in the Pohatcong Creek drainage basin, on Brass Castle Creek near Washington, and Pohatcong Creek at New Village. The locations of the continuous-record stations are shown in figure 9, and results of hydrograph-separation analyses are given in table 4.

Results of the base-flow separation analyses indicate that annual average base flow of Brass Castle Creek near Washington, N.J., (01455160) was 2.6 ft^3/s (73 percent) during water years[1] 1978-83. Annual average base flow of Pohatcong Creek at New Village (01455200) was 19.1 ft^3/s during water years 1960-69, about 72 percent of total flow. However, because drought conditions occurred during 1961-65, average annual flow and base flow for Pohatcong Creek at New Village (01455200) during the period of record were lower than average. The period of record for Brass Castle Creek near Washington (01455160), 1978-1983, included both years of drought

[1] A water year is the 12-month period from October 1 through September 30 and is designated by the calendar year in which it ends.

Table 4. Summary of discharge and base flow (determined using base-flow separation techniques) for continuous-record stream-gaging stations in and near the Pohatcong Valley study area, Warren County, New Jersey.

[ft^3/s, cubic feet per second]

Flow-gaging station number	Flow-gaging station name	Drainage area (square miles)	Period of record	Years of daily discharge record	Mean annual discharge[1] (ft^3/s)	Mean annual base flow[2]	
						in ft^3/s	in percent of total flow
01440000	Flat Brook near Flatbrookville, N.J.	64.0	1923-2000	78	111	81.8	74.0
01443500	Paulins Kill at at Blairstown, N.J.	126	1921-2000	80	197	155	77.4
01445500	Pequest River at Pequest, N.J.	106	1921-2000	80	157	131	83.4
[3]01455160	Brass Castle Creek near Washington, N.J.	2.34	1978-83	5	3.6	2.6	73.0
[3]01455200	Pohatcong Creek at New Village, N.J.	33.3	1959-2000	10	26.6	19.1	71.6
01457000	Musconetcong Riv. near Bloomsbury, N.J	141	1903-07, 1921-2000	81	239	203	84.9

[1]Values from Reed and others (2001).

[2]Base flow was estimated by using the methodology of Rutledge (1993) with data for the water year. The water year is the 12-month period from October 1 through September 30. It is designated by the calendar year in which it ends.

[3]Station also was used as a low-flow partial-record station because of the short period of record.

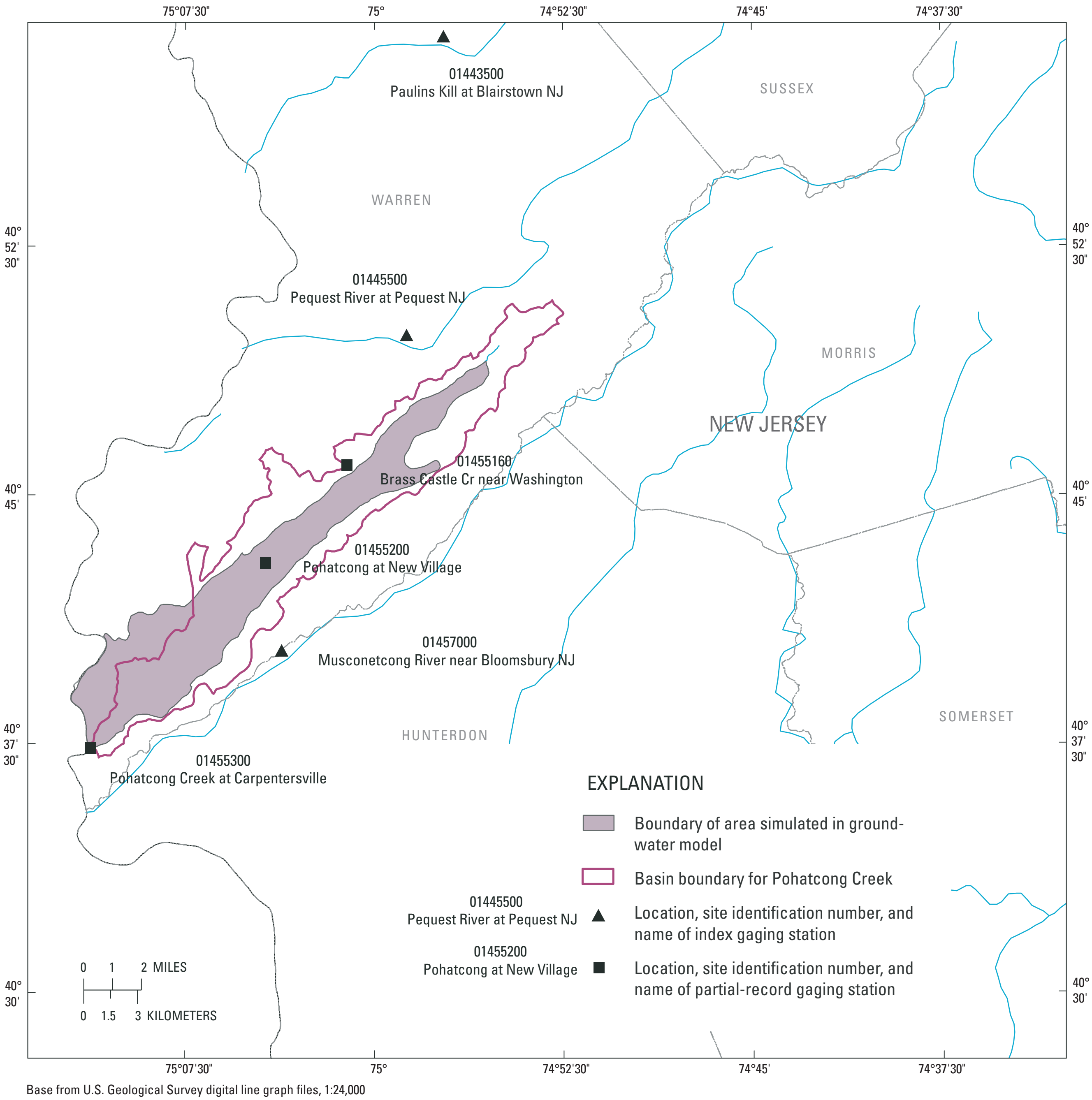


Figure 9. Location of index and partial-record streamflow stations in the vicinity of the Pohatcong Valley, Warren County, N.J.

and years of unusually high precipitation, and the mean annual flow estimate is thought to be higher than the long-term average, on the basis of estimates from low-flow correlations (discussed in the following section). The short duration of record for the stations at Brass Castle Creek and Pohatcong Creek and the extreme conditions that occurred during both periods of record render the results of base flow-separation analysis for these data somewhat unreliable; therefore, additional analyses were made by correlating individual measurements of discharge at these two sites and at one other site within the study area with discharge data from four index stations as described below.

Estimated Mean Annual Base Flow at Partial-Record Stations

The mean annual discharge and mean annual base flow at stations for which continuous-record data are unavailable can be estimated by correlating instantaneous low-flow discharge at a low-flow, partial-record station with the concurrent mean daily discharge at a continuous-record index station in a similar hydrogeologic setting (Gillespie and Schopp, 1982). In the Pohatcong valley study area, three low-flow partial-record stations were selected for use in the low-flow correlation analyses-- Brass Castle Creek near Washington, Pohatcong Creek at New Village, and Pohatcong Creek at Carpentersville.

Values of measured discharge at each of these low-flow, partial-record stations were correlated with mean daily discharge at four index gaging stations adjacent to the study area-- Pequest River at Pequest, the Musconetcong River near Bloomsbury, Flat Brook near Flatbrookville, and Paulins Kill at Blairstown (fig.9). The periods of continuous record at the two stream gaging stations within the study area, Brass Castle Creek near Washington and Pohatcong Creek at New Village, are of short duration; therefore, estimates of mean annual base flow derived on the basis of correlations with low-flow measurements made over a longer period are considered more accurate. The low-flow correlation equation is used to estimate, or predict, discharge statistics at the low-flow partial-record station, QP_R, on the basis of the values of the same discharge statistics measured at the index gaging station, QI. The correlation coefficient is an indication of the accuracy of the predicted discharge. The correlation coefficient is a number from –1.0 to 1.0 that indicates the strength of the linear relation between the logarithm (base 10) of the discharge at the low-flow partial record station and that at the index station. The nearer the correlation coefficient is to +/-1.0, the more reliable is the predicted discharge, QP_R.

The estimated mean annual flow and mean annual base flow at the four index stations were input to the correlation equations to determine estimated mean annual flow and mean annual base flow at each of the low-flow partial-record stations. The low-flow partial-record stations and their associated correlation equations are listed in table 5. The average of the predicted mean annual flows for Brass Castle Creek near Washington is 2.32 ft^3/s (14 in/yr over the basin area), and the average of the predicted mean annual base flows is 1.7 ft^3/s (10 in/yr), 73 percent of mean annual flow. The average of the predicted mean annual flows for Pohatcong Creek at New Village is 32 ft^3/s (13 in/yr), and the average of the predicted mean annual base flows is 25 ft^3/s (10 in/yr, 78 percent of mean annual flow. The average of the predicted mean annual flows for Pohatcong Creek at Carpentersville is 54 ft^3/s (13 in/yr), and the average of the predicted mean annual base flows is 45 ft^3/s (11 in/yr), 82 percent of mean annual flow. Estimated base flow for Pohatcong Creek at Carpentersville is about 1 in/yr greater than that estimated for Pohatcong Creek at New Village, although the estimated flow was about the same. The difference in the base flow could be due to the inherent uncertainties in the estimates, greater discharge from the bedrock aquifer to Pohatcong Creek by way of the surficial aquifer, or other unknown differences between the contributing areas. Direct runoff (mean annual flow – base flow) is about 3.5, 3.0, and 2.3 in/yr at the Brass Castle Creek, Pohatcong Creek at New Village, and Pohatcong Creek at Carpentersville, respectively. The higher value of direct runoff in Brass Castle Creek is most likely a consequence of the higher percentage of steep slopes in this basin than the other basins. In addition, the Brass Castle Creek basin is underlain by poorly permeable crystalline bedrock, whereas the Pohatcong Creek basin includes parts of the valley that have gentler slopes and thicker unconsolidated sediments, which are underlain by the more permeable carbonate bedrock.

Direct Discharge to the Delaware River from the Carbonate-Rock Aquifer

The rate of discharge from the carbonate-rock aquifer directly to the Delaware River (therefore, bypassing the local ground-water-discharge-to-streams system) is not known. An approximation of instantaneous flux in the carbonate-rock aquifer can be calculated using Darcy's Law, where discharge is equal to hydraulic conductivity times the cross-sectional area of flow times the head gradient. Where the thickness of the aquifer is not well determined, the flux can be calculated by multiplying transmissivity times the width of the valley times the gradient. It is difficult to estimate the width of the aquifer where it intersects the Delaware River (at Carpentersville) because carbonate rocks from the Lopatcong Valley to the north merge with the carbonate rocks of the Pohatcong Valley. Furthermore, no ground-water head gradient data are available for the Carpentersville area. Therefore, to estimate aquifer discharge, the valley width and aquifer head gradient were estimated at New Village. The valley is about 6,500 ft wide at New Village, and the head gradient in the carbonate-rock aquifer is about 0.0035 ft/ft. Using transmissivity estimates ranging from 2,500 to 25,000 ft^2/d, downvalley flux in the carbonate-rock aquifer at New Village range from 0.66 to 6.6 ft^3/s . Converting discharges to inches per year using the basin area of Pohatcong Creek at New Village of 33.3 mi^2, flux estimates range from 0.27to 2.7 inches per year.

Table 5. Correlation equations relating instantaneous low-flow measurements at low flow partial-record gaging stations to concurrent mean daily flow at continuous-record stream-gaging stations (index stations) in the Pohatcong Valley study area, New Jersey

[Station locations shown in figures 8 and 9. Q_{PR}, predicted discharge at partial-record station; Q_I, discharge at index station; N.J., New Jersey.]

Low-flow partial-record-gaging station number	Low-flow partial-record-gaging station name	Drainage area (square miles)	Index gaging-station number	Number of measurements used in analysis	Years of daily discharge record	Correlation coefficient[1]	Correlation equation	Predicted mean annual discharge (Q_{PR})[2] (cubic feet per second)	Predicted mean annual base flow[3] (Q_{PR}): Cubic feet per second	Predicted mean annual base flow[3] (Q_{PR}): Inches per year over the drainage area	Predicted mean annual base flow[3] (Q_{PR}): Percent of total flow[4]
[5]01455160	Brass Castle Creek near Washington, N.J	2.34	01445500	99	80	0.88	$Q_{PR} = 0.00104\,Q_I^{(1.5251)}$	2.0	1.8	10	78.3
			01457000	100	81	.87	$Q_{PR} = 0.00011\,Q_I^{(1.7864)}$	2.0	1.5	8.5	75.0
			01443500	92	80	.85	$Q_{PR} = 0.00119\,Q_I^{(1.4305)}$	2.3	1.6	9.4	69.6
			01440000	96	78	.83	$Q_{PR} = 0.00616\,Q_I^{(1.3132)}$	3.0	2.0	12	66.7
[5]01455200	Pohatcong, Creek at New Village, N.J.	33.3	01445500	47	80	.94	$Q_{PR} = 0.06985\,Q_I^{(1.2151)}$	33	26	11	78.8
			01457000	45	81	.89	$Q_{PR} = 0.01149\,Q_I^{(1.4268)}$	28	23	9.2	82.1
			01440000	48	78	.87	$Q_{PR} = 0.26992\,Q_I^{(1.026)}$	34	25	10	73.5
			01443500	45	80	.86	$Q_{PR} = 0.08343\,Q_I^{(1.137)}$	34	26	11	76.5
01455300	Pohatcong Creek at Carpentersville, N.J.	57.0	01457000	18	81	.95	$Q_{PR} = 0.11144\,Q_I^{(1.1289)}$	54	45	11	83.3
			01443500	17	80	.94	$Q_{PR} = 0.71547\,Q_I^{(0.8134)}$	53	43	10	81.1
			01445500	19	80	.94	$Q_{PR} = 0.6352\,Q_I^{(0.8667)}$	51	43	10	84.3
			01440000	17	78	.93	$Q_{PR} = 1.62863\,Q_I^{(0.7618)}$	59	47	11	79.7

[1]Index-gaging stations are given in order of decreasing correlation coefficient for each low-flow partial-record gaging station.

[2]Values calculated using mean annual discharge for index station reported in Reed and others (2001).

[3]Base flow was estimated by using the methodology of Rutledge (1993) using data for the water year. The water year is the 12-month period from October 1 through September 30. It is designated by the calendar year in which it ends. For baseflow in this study, the periods of record begin with the earliest October 1 and end with the latest September 30.

[4]Percentage of total flow = Predicted mean annual base flow divided by mean annual QP_R ,in cubic feet per second.

[5]Station was also a continous-record stream-gaging station.

Ground- and Surface-Water Withdrawals

Large-scale (greater than 10 Mgal/yr) ground- or surface-water withdrawals are made at only a few sites in the Pohatcong Valley. The only substantial withdrawals are from production wells PVMSW01 and PVMSW04 (and PVMSW02 in the adjacent Musconetcong Valley), and four wells at industrial facilities in Washington Borough (P484, P485, P494, and P495). Wells P494 and P495 are open to the crystalline-rock aquifer and, therefore, are not directly in the flow system modeled for this study. The water withdrawn from wells P484 and P485 is used for non-contact cooling, and about 95 percent of the water withdrawn is re-injected into the carbonate-rock aquifer through well P483 (table 6). An average of about 2 Mgal/yr was withdrawn from well C493 in Washington Borough for cooling purposes and from Pohatcong Creek and a pond for irrigation of a golf course. Two farmers in the study area have surface-water withdrawal permits but did not report any withdrawals from 1997 to 2002. Some irrigation ponds have been observed in the valley, but the water quantities withdrawn from these ponds, if any, are not known. Agricultural withdrawal quantities are believed to be small and are considered negligible here. In addition to the wells with reported withdrawals, hundreds of domestic wells are present in the valley, primarily north of Washington Borough, along the northern edge of the valley, and southwest of the village of Broadway. Wells PVMSW01, PVMSW04, P483, P484, P485, and C493 are included in the numerical model, as are all domestic wells within the area underlain by the carbonate-rock aquifer for which locations are known.

Ground-Water Levels

Ground-water levels were measured in 2001 and 2002 to determine vertical gradients, horizontal gradients, and estimate the direction(s) of ground-water flow (Carleton and others, 2005; CH2M Hill, 2005a). Land-surface altitudes for about 100 of the wells were measured with 0.3-ft accuracy or better. Because of a pervasive drought from September 2001 to March 2002, water levels were lower in June 2002 than in June 2001. For 75 wells measured in both years, the median difference, mean difference, and standard deviation were 3.49 ft, 4.26 ft, and 4.76 ft, respectively, and 2002 water levels were at least 1 ft lower than 2001 levels in about 85 percent of the wells. Therefore, the 2001 synoptic water-level data are considered more representative of long-term average conditions than the 2002 synoptic data (figs. 10 and 11).

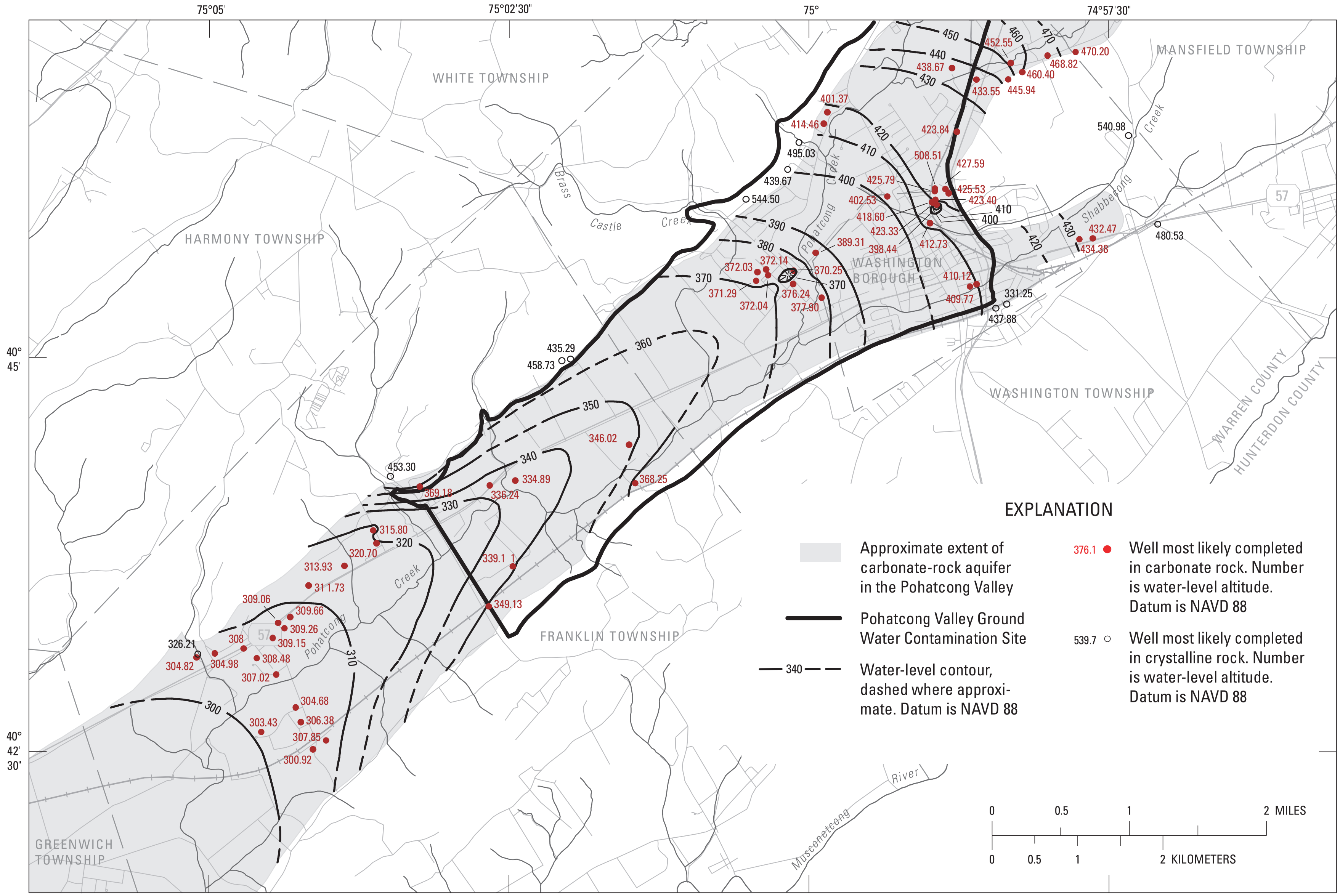
In some locations, water levels in the shallow part of the surficial aquifer (in wells screened 5-20 ft below the first detected saturated zone) were as much as 63 ft higher than in

Table 6. Ground- and surface-water withdrawals, Pohatcong Valley, Warren County, N.J.

[NJDEP, New Jersey Department of Environmental Protection; BWA, Bureau of Water Allocation; Mgal/yr, million gallons per year; -- not available; NA, not applicable]

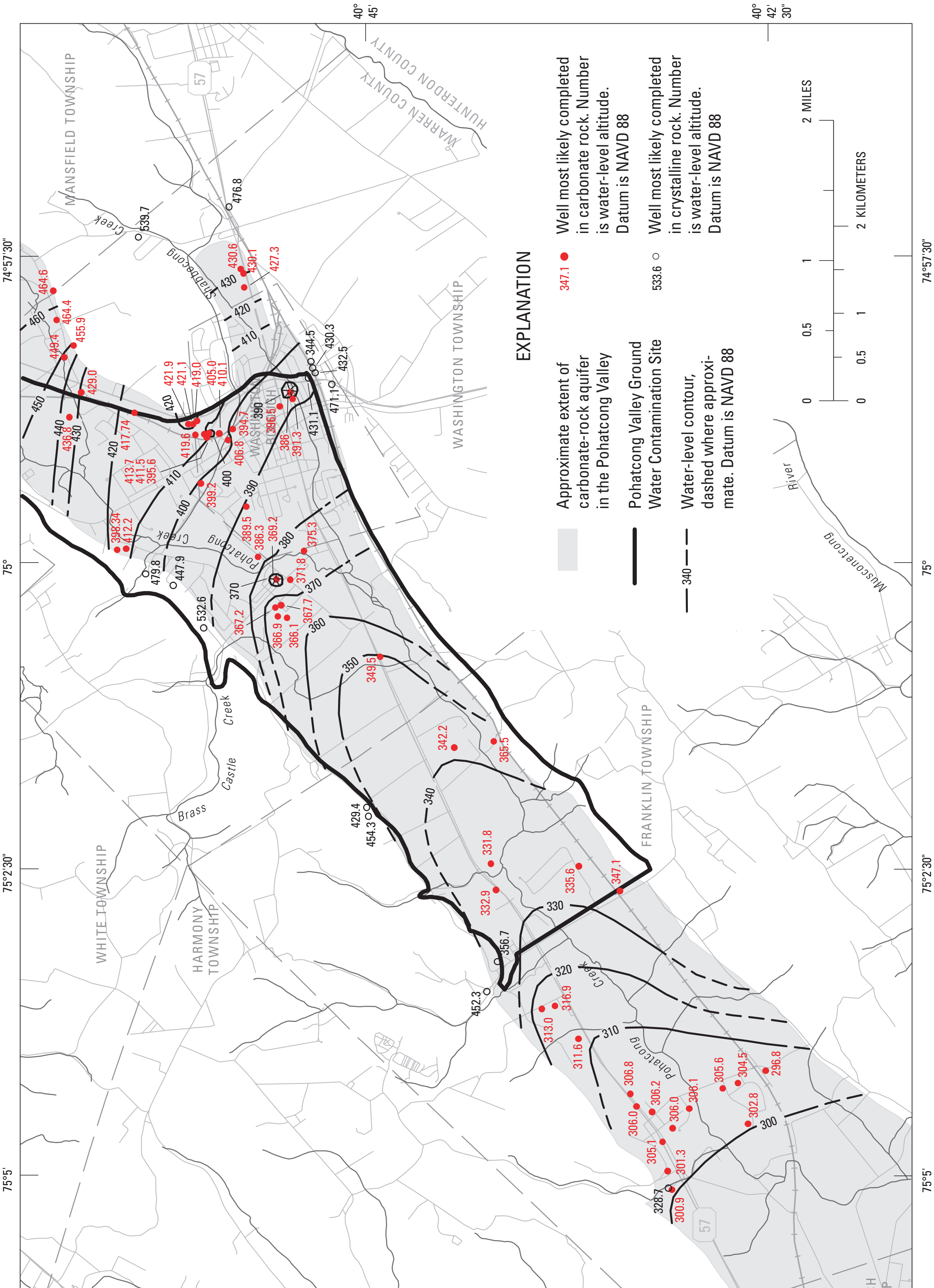
Facility type	NJDEP BWA permit number	NJDEP well permit number	Local identifier	Average reported annual withdrawals, 1997-2002 (Mgal/yr)	Aquifer type	Included in numerical model
Public supply	5053	24-8261	PVMSW01	117	Carbonate rock	Yes
		24-12183	PVMSW04	58	Carbonate rock	Yes
		24-16653	PVMSW02	133	Carbonate rock	No
Industrial	2050P	24-4975,	P484	-108[1]	Carbonate rock	Yes
		24-22618	P485			
		--	P483			
Industrial	10429W	24-22566	P494	15	Crystalline rock	No
		24-31397	P495			
Commercial	10781W	24-3250	C493	2	Carbonate rock	Yes
Golf course	10705W	--	Pohatcong Creek, Pond 1	2	NA	No
Farm	WA0022	--	Pond	0	NA	No
Farm	WA041R	--	Pond	0	NA	No

[1] Negative value indicates water was injected into the aquifer.



Base from U.S. Geological Survey digital line graph files, 1:24,000

Figure 10. Water levels in wells completed in the carbonate-rock aquifer, and water-level contours, Pohatcong Valley, Warren County, N.J., June 2001.



Base from U.S. Geological Survey digital line graph files, 1:24,000

Figure 11. Water levels in wells completed in the carbonate-rock aquifer, and water-level contours, Pohatcong Valley, Warren County, N.J., June 2002.

the deep part of the aquifer (5-20 ft above the top of bedrock) and in the underlying carbonate-rock aquifer. Water-level altitudes in June 2002 in the shallow part of the surficial aquifer and in streams were higher than in the underlying carbonate-rock aquifer in all locations. Water-level altitudes in June 2001 in the carbonate-rock aquifer were higher than those in the adjacent Pohatcong Creek in the upper part of the valley (northeast of Washington Borough) but were lower than surface-water altitudes in the central part of the valley. In July 2000, the water-level altitude in one bedrock well (PVWCC15) was 0.1 ft higher than the altitude of nearby Pohatcong Creek, but the water level in a second bedrock well (PVFCC12) closer to Washington Borough was lower than that in nearby Pohatcong Creek. The upward and downward water-level gradients measured at different times indicate an upward vertical gradient from the carbonate-rock aquifer to the surficial aquifer and overlying Pohatcong Creek in most locations near the creek during periods of high water levels, and a downward gradient in most locations throughout the valley during periods of low water levels. The average condition is not known, but a slight upward gradient (less than 0.5 ft) probably occurs directly under Pohatcong Creek and a downward gradient occurs elsewhere. At the downstream end of the Pohatcong Valley, ground water most likely discharges from the carbonate-rock aquifer directly to the Delaware River.

Vertical gradients differ throughout the valley, apparently a consequence of the variable hydraulic properties of the surficial aquifer. In the vicinity of well PVMSW01, water-level altitudes in the shallow part of the surficial aquifer and in Shabbecong Creek were about 40 ft higher than those in the bedrock. Water-level altitudes in June 2002 in wells D437, A3, and D412 (north of Washington Borough) were about 2.0, 1.4, and 0.6 ft lower, respectively, than in adjacent Pohatcong Creek or in well D411 (completed in the shallow part of the surficial aquifer close to Pohatcong Creek). Farther downstream, water-level altitudes in June 2002 in observation wells PVFCC12, PVBMC02, and PVWCC15 and in domestic well D225 were about 8, 28, 4, and 3 ft lower, respectively, than nearby surface-water altitudes.

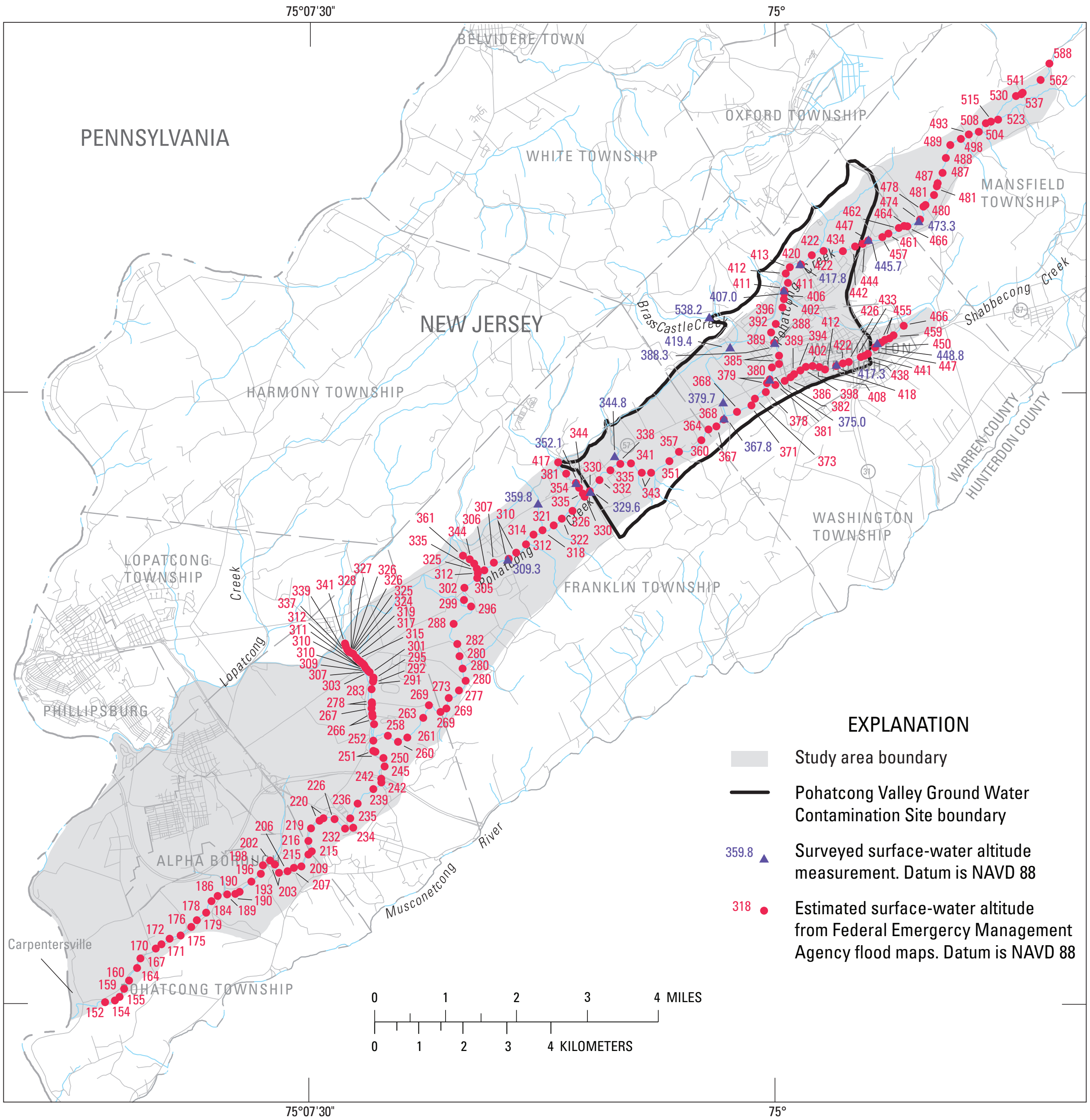
Surface-water altitudes at 22 locations were surveyed to within 0.03 ft (CH2M Hill, 2005a). Additional surface-water altitudes for Pohatcong Creek, Shabbecong Creek, Brass Castle Creek, and other tributaries within the study area were estimated from flood profile cross-sections included in Federal Emergency Management Agency (FEMA) flood-insurance studies (Federal Emergency Management Agency, 1982a, 1982b, 1982c, 1982d, 1983). The average depth of Pohatcong Creek, based on six measurements made along the length of the creek within the study area, is about 1 foot. Therefore, the surface-water altitudes for Pohatcong Creek were generated by adding 1 ft to the streambed altitudes reported in FEMA flood-insurance studies at each cross section. In a few locations, a dam or other restriction caused the creek to be deeper than average. At such locations, more than 1 ft was added to the streambed altitude to estimate the surface-water altitude. For Shabbecong Creek, Brass Castle Creek, and other tributaries, 0.5 ft was added to the streambed altitudes. Eight of 14 surface-water altitudes estimated from flood studies at or near one of the surveyed surface-water sites are within 0.7 ft of the surveyed altitude. At least three of the sites with a difference between estimated and surveyed water-level altitudes greater than 0.7 feet are in areas of steep gradient or in areas where recent modifications to the channel could explain the larger difference. Estimated and surveyed surface-water altitudes are shown in figure 12.

Ground-Water Age

Water samples collected from 2 wells screened in the surficial aquifer and 10 wells open to the carbonate-rock aquifer were analyzed to estimate the age of the water. Concentrations of sulfur hexafluoride (SF_6) were analyzed in all 12 samples, chlorofluorocarbons (CFCs) in 8 samples, and tritium-helium in 4 samples. Relatively inert gases can be used to estimate when ground water entered the aquifer (and was last in contact with the atmosphere) if the concentration of the gas in the atmosphere over time is known and the concentration is monotonically increasing or decreasing. Measuring the concentration of the gas in the water and comparing the known solubility of the gas allows the concentration of the gas in the atmosphere when the water last contacted the atmosphere to be calculated. The year that the atmosphere contained the determined concentration of the inert gas is the year in which the sampled water entered the ground-water system. The technique is most effective when the atmospheric concentration is changing monotonically and rapidly, so that the changes in concentration over short periods are measurable and represent a definable period. Target gas concentrations that exceed atmospheric levels (indicating a local source not correlated to average northern hemisphere concentrations) invalidate the age-dating techniques.

CFCs have been used for age-dating water for more than a decade (Busenberg and Plummer, 1991; Busenberg and Plummer, 1992; Pope and others, 1999; Szabo and others, 1996). Atmospheric concentrations began to increase rapidly beginning in the 1930s and concentrations of most CFCs began to stabilize in the mid-1990s.The atmospheric concentrations of some CFCs are now decreasing because of international agreements to cease production in 1996 (Plummer and Friedman, 1999), which means that for some concentrations in ground water, that water could be one of two different ages. Furthermore, in industrial areas CFCs commonly are present in much higher concentrations than the northern hemisphere average. Water that has recharged the aquifer in an area where higher atmospheric concentrations are present locally can have concentrations greater than possible from globally averaged atmospheric concentrations, and ages cannot be determined.

SF_6 is a heavy, inert gas, the atmospheric concentration of which is monotonically increasing very rapidly. The concentrations are about two orders of magnitude less than those of the CFCs, but the age of water can still be determined to within a narrow range. Because concentrations are still



Base from U.S. Geological Survey digital line graph files, 1:24,000

Figure 12. Surveyed and estimated surface-water altitudes on Pohatcong Creek and selected tributaries, Warren County, N.J.

increasing rapidly and fewer local sources of contamination exist, SF_6 is often more effective for dating water that has been in contact with the atmosphere after about 1975. Busenberg and Plummer (USGS, written commun., 2000) report that there are rare terrigenic sources of SF_6 and that Silurian-age carbonates can have relatively high naturally occurring concentrations of SF_6.

Tritium is a naturally occurring, relatively short-lived radio isotope (half-life about 12.3 years) that was introduced into the atmosphere at concentrations far exceeding natural concentrations by atmospheric testing of thermonuclear devices. Therefore, tritium can be used as a qualitative tracer to determine whether the water was in contact with the atmosphere prior to, or after the onset of, atmospheric testing in 1953. The atmospheric concentrations peaked in 1963, then began decreasing rapidly because of the ban on atmospheric testing. The concentration of tritium and its decay product, tritiogenic helium, can be used as a conservative tracer and, by their ratio, indicate the length of time of radioactive decay (the time since tritium was isolated from the atmosphere).

Ages of water estimated from SF_6 concentrations ranged from modern (0-2 years) to 27 years, with a median of 6 years (table 8, farther on in the report, fig. 13). Invalid concentrations precluded age determination using CFCs in all but two of these samples, one of which contained SF_6 concentrations exceeding atmospheric levels. The estimated ages from CFC concentrations (0-2 years) and SF_6 concentration (3 years) are similar in the one well (PVWCC15) for which apparently valid samples were analyzed using both techniques. The median age determined from all valid SF_6 and CFC samples is 6 years. In three of the four wells with ages determined using tritium-helium, the SF_6 age was slightly more than half the tritium-helium age. In the fourth well, the SF_6 age, 27 years, was more than twice any other SF_6 age; yet the tritium-helium age was 6 years. The reason for this difference is not known.

The ages determined for water samples from adjacent wells PVFCC12 and PVFCC13 (completed in carbonate rock and overlying unconsolidated sediments, respectively) differ by a factor of three: the estimated age of water from the bedrock well is three times that of water from the unconsolidated well. Results for the other sampled well pair, PVWCC14 and PVWCC15, are ambiguous, in part because the samples are thought to be contaminated by a local source, invalidating the technique. An increase in age with depth is expected in a recharge area (Szabo and others, 1996).

Mixing of waters of different ages limits the application and reliability of the technique of age-dating ground water by measuring concentrations of dissolved atmospheric gases. Furthermore, in fractured-rock settings, gases may diffuse into the low-permeability rock matrix, resulting in a loss of the target gas. In the wells adjacent to or penetrating fractured-rock sequences, the sample might represent a mixture of waters with different residence times–in the simplest case, a mixture of younger and older waters. The use of multiple tracers in the absence of mixing ought to result in a series of internally consistent residence times (Szabo and others, 1996). If mixing has occurred, ratios of the tracers can be used to approximate the apparent age of the younger fraction of the water (Plummer and others, 1998; Plummer and others, 2001). For this study, invalid CFC concentrations precluded the analysis of mixed ground waters with different ages.

Estimated ages of the carbonate-rock-aquifer water are in a narrower range (0 to 27 years) than, and do not increase in age in the downstream direction as indicated, in other settings (Szabo and others, 1996; Modica, 1996). Water-level contours (figs. 10 and 11) indicate that the direction of flow in the carbonate-rock aquifer is downvalley (southwest); therefore, ground-water ages should increase in the downvalley direction. The ages shown in figure 13 do not indicate an increase in age in the downvalley direction, most likely because any such variation is less than the variation in the results caused by other factors. Complicating factors, including anthropogenic contamination, mixing of different-age waters in the heterogeneous, fractured-rock aquifer, and mixing within the wells from which the samples were collected, obscure the small age differences in the relatively young water.

The relatively young water (less than 27 years) clearly indicates that ground water moves quickly through the carbonate formations. The young age is important for this study because it indicates that, with respect to advection, the contaminant plumes identified by CH2M Hill (2005a) could have reached steady state because the sources are assumed to be older than even the oldest ground-water age recorded. (Contamination was first identified in the valley in 1978 in concentrations similar to concentrations measured about 25 years later.) Processes other than advection, however, including natural attenuation, could alter contamination plumes.

Water Budgets

Two system-wide water budgets for the aquifer system were developed; a land-surface water budget and ground-water budget (Nicholson and others, 1996, p. 66). The land-surface budget accounts for precipitation (the input to the system) that falls on the land surface and was used to estimate the rate of direct recharge to the surficial aquifer within the Pohatcong Valley. The ground-water budget accounts for water entering the ground-water system through all sources of recharge (inputs) and leaving the system as discharge (outputs). Precipitation that falls on areas underlain by valley-fill deposits and on the surrounding upland areas infiltrates to the ground-water flow system and is the principal source of recharge to the aquifers. Results of studies of several areas in the glaciated northeastern United States indicate that an appreciable percentage of the natural recharge to glacial valley aquifers is derived from upland runoff (Morrissey and others, 1987). Recharge from upland areas includes seepage losses from upland-draining tributaries where those tributaries enter the valley, infiltration of unchanneled runoff at the bases of hillsides, and underflow of ground water from till or bedrock.

The land-surface budget for valley areas is represented by the equation:

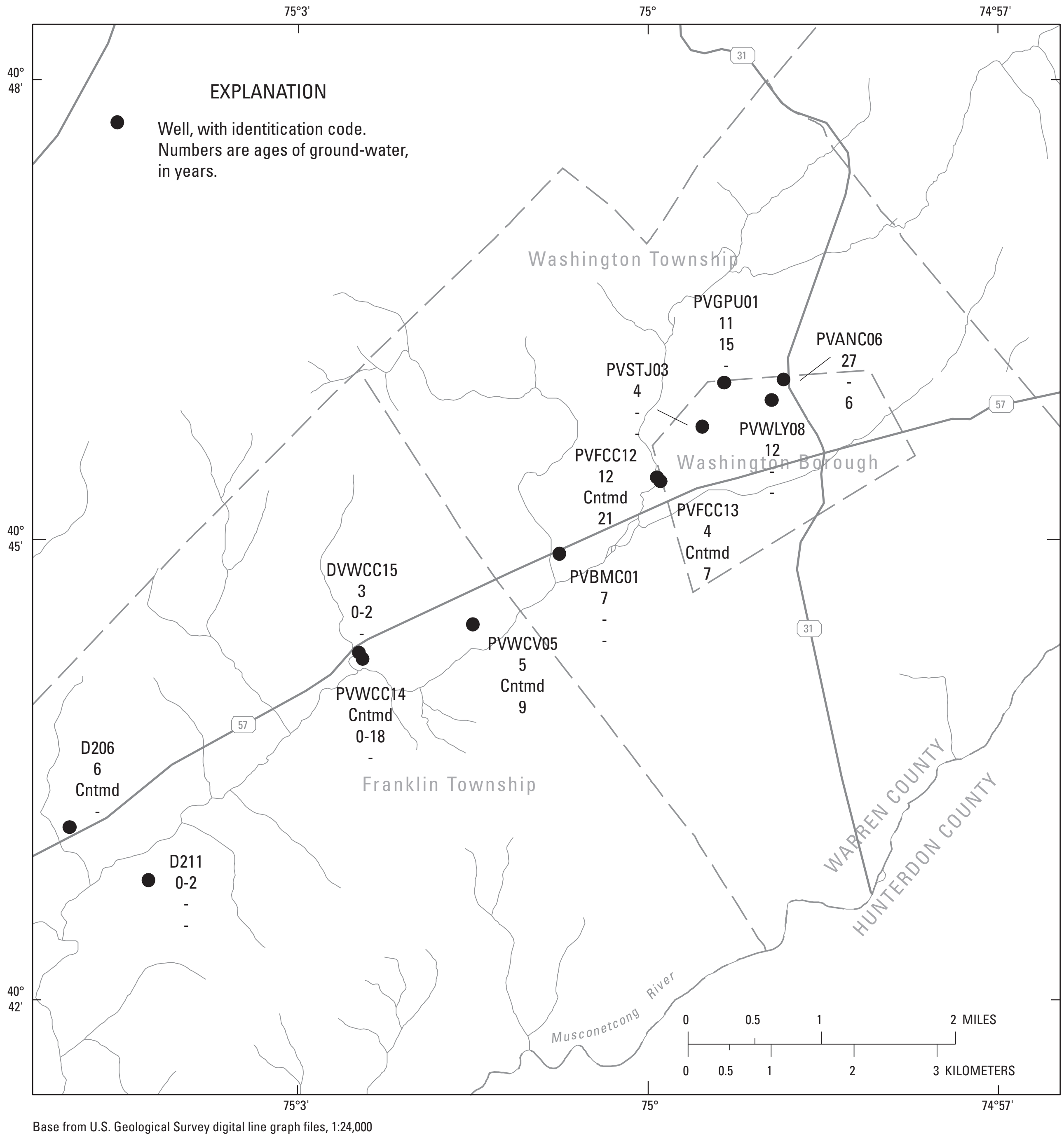


Figure 13. Location of selected wells for which ground-water ages were calculated from sulfur hexafluoride (SF6), chlorofluorocarbon (CFC), and tritium-helium isotope (H^3-He^3) concentrations, Pohatcong Valley, Warren County, N.J. (First number is from SF_6 method; second number, CFC method; third number, H^3 - He^3. Cntmd indicates age could not be determined because sample contained SF_6 or CFC contamination. - indicates no analysis for CFC or He^3 - He^3 method.)

$$P = ET + DR + R, \qquad (1)$$

where

P = precipitation,
ET = evapotranspiration,
DR = direct runoff,

and

R = direct recharge to the valley aquife system.

The average precipitation during 1971-2000 at three sites (Belvidere Bridge, Long Valley, and Split Rock Pond) was 51 in/year (Robinson, 2003). Using the Thornthwaite method (Thornthwaite and Mather, 1957, as adapted by Mather, 1978), Nicholson and others (1996, p. 67) estimated potential evapotranspiration in the region to be about 25 in/yr. No estimate of direct runoff in a stream in the Pohatcong Valley underlain only by carbonate rocks is available. Direct runoff in Pohatcong Creek at Carpentersville (discussed in a preceding section on base flow) is about 2.3 in/yr, compared with 3.5 in/yr for the part of Brass Castle Creek underlain entirely by crystalline rock. Therefore, an estimate of 3 in/yr is considered accurate enough for the current exercise. Using the above estimates for precipitation, evapotranspiration, and direct runoff, direct recharge to the valley aquifer system is estimated to be about 23 in/yr.

The ground-water budget for the valley floor (area underlain by carbonate rock) aquifer system is defined by

$$R = B_P + B_D + P - W - T, \qquad (2)$$

where

R = total recharge to the aquifer system,
B_P = ground-water discharge to and treated wastewater return flow for Pohatcong Creek (base flow),
B_D = ground-water discharge directly to the Delaware River,
P = net ground-water pumpage (withdrawal – injection),
W = treated wastewater returns,

and

T = throughflow (sum of base flow of upland tributaries entering valley, less streamflow loss to the aquifer system from upland tributaries).

The recharge term (R) for the valley aquifers consists of direct infiltration of precipitation, infiltration of streamflow from upland tributaries, and infiltration of runoff from unchanneled upland areas (fig. 3). Because direct infiltration of precipitation and runoff from unchanneled upland areas is difficult to measure, recharge commonly is estimated by looking at the outputs of the aquifer system, in this case, stream base flow and ground-water withdrawals. Base flow in Pohatcong Creek consists of more than discharge from the carbonate-rock and surficial aquifers: effluent from the Washington sewage treatment plant and base flow in tributaries entering the valley floor (carbonate-rock area) from the valley walls (crystalline-rock area) augment the flow in the creek. Therefore, these terms must be subtracted from the base flow in the creek to determine ground-water discharge from the carbonate-rock aquifer to the creek.

The budget for the part of the basin upstream from New Village was estimated because few data were available for the entire basin. Estimates of flow rates such as pumpage (P) and wastewater returns (W) were divided by the area of the basin underlain by carbonate rock upstream from New Village, 11.4 mi^2. The total annual pumpage (P) from the aquifer system for 1997-2001 was 0.76 Mgal/d (1.4 in/yr). The average annual wastewater return (W) for 2000 from the Washington Borough Sewage Treatment Plant was reported to be 0.71 Mgal/d (1.3 in/yr) (New Jersey Department of Environmental Protection, 2005). The estimated average annual base flow for Pohatcong Creek at New Village is 25 ft^3/s (30 in/yr). In a previous section, the mean annual base flow for Pohatcong Creek at New Village of 25 ft^3/s was converted to 10 in/yr. The difference arises because the area of the basin underlain by crystalline and carbonate rock, 33.3 mi^2, was used in the previous section, whereas in this section only the carbonate-rock area of the basin, 11.4 mi^2, was used to calculate the conversion.

Throughflow is that part of base flow in Pohatcong Creek at New Village that originates in the upland (crystalline-rock) area, emerges in tributaries in that part of the valley underlain by carbonate rock, and exits the valley in Pohatcong Creek without contacting ground water in the carbonate rock. The average annual base flows for streams draining upland areas underlain by crystalline rocks generally are not known. The base flows were measured, however, in almost all tributaries entering Pohatcong Creek upstream from New Village in August 2000. The measured flows in tributaries, including Brass Castle Creek at the confluence with Pohatcong Creek, were summed and the result (6.6 ft^3/s) multiplied by the ratio of estimated mean annual base flow of Pohatcong Creek at New Village (01455200) to the measured flow on August 8, 2000, 25 ft^3/s to 20.9 ft^3/s, to make a crude estimate of mean annual base flow on the tributaries, 7.9 ft^3/s (9.4 in/yr). This estimate does not include throughflow entering Pohatcong Creek above the Tunnel Hill Rd station (01455135). No data are available for the volume of throughflow. To estimate a maximum possible throughflow, the discharge at that station, 6.78 ft^3/s, was multiplied by the 25/20.9 ratio and added to the previously estimated throughflow. The resulting estimate, 16 ft^3/s (19 in/yr) represents a likely maximum value for throughflow. This analysis also does not account for losses to the aquifer from the measured tributaries between the gaging site and the confluence with Pohatcong Creek, but those losses are believed to be small.

Discharge from the carbonate-rock aquifer to the Delaware River (B_D) is difficult to measure. Thus discharge was estimated using (also) estimated values for streambed conductance, aquifer transmissivity, and hydraulic gradient (discussed in the “Direct discharge to the Delaware River from the carbonate-rock aquifer" section of this report). The estimates of flow in the carbonate-rock aquifer at New Village range from 0.66 to 6.6 ft^3/s, or 0.79 to 7.9 in/yr over the 11.4 mi^2 carbonate-rock part of the basin. The highest value is believed to be the most likely; therefore, 7.9 in/yr was used in water-budget calculations. (To compare to units used in previous sections, convert flows to inches per year using the basin area of Pohatcong Creek at New Village of 33.3 mi^2: flow estimates range from 0.27 to 2.7 inches per year.)

Substituting the estimated values above that are considered to be most accurate into each term on the right side of equation (2) yields

$$R = 30 + 7.9 + 1.4 - 1.3 - 9.4 = 28.6 \text{ in/yr}$$

These results are similar to those used in a calibrated model of ground-water-flow in the New Jersey Highlands in Nicholson and others (1996), which also simulated flow in a glacial overburden and carbonate-rock aquifer system in southwestern Morris and northeastern Hunterdon Counties. Nicholson and others (1996) estimate that direct recharge to the valley aquifers is 22 in/yr, plus 6 in/yr (14 ft^3/s) of infiltrating runoff from adjacent, unchanneled upland areas and 2.6 in/yr (5.9 ft^3/s) of tributary leakage to the aquifer system, for a total of 30.6 in/yr. Because estimates for the throughflow (T) and direct aquifer discharge to the Delaware River (B_D) terms in the ground-water budget equation above are subject to considerable error, the recharge (R) was refined during calibration of the ground-water-flow model.

Simulation of Ground-Water Flow

A numerical model was constructed to simulate ground-water flow in the Pohatcong Valley. The numerical model is based on the conceptual model of flow described below, which in turn was based on the geology and hydrogeology of the valley described in preceding sections of this report. The input values (parameters) of the model were taken from results of the investigations described in the preceding sections and were modified during calibration when those values were estimated or uncertain.

Conceptual Model of Ground-Water Flow

The highly transmissive carbonate formations underlying the Pohatcong Valley constitute the major aquifer within the study area. The lithology between, and even within, the individual carbonate formations varies, but there are no known thick and widely present units (such as the Martinsburg Shale found nearby) within the valley that act as major confining units. Outcrops of upper Allentown Dolomite and Beekmantown Group along the Delaware River at the southwestern end of the Pohatcong valley show these rocks have similar lithology and fracturing, with some variation from massively-bedded units to more finely bedded, fissile units. The five carbonate formations are considered to be a single aquifer. Aquifer test results indicate the carbonate-rock aquifer has a high horizontal anisotropy, with horizontal hydraulic conductivity along strike (approximately parallel to the axis of the valley) greater than that perpendicular to strike (perpendicular to bedding planes).

The unconsolidated sediments overlying the carbonate-rock aquifer form a complex unit; in some locations they (the sediments) act as an aquifer, and in other locations they apparently are unsaturated to depths of up to 110 ft. Perched water tables have been identified in several locations, including the vicinity of the PVANC wells and the production well PVMSW04. Elsewhere a significant saturated thickness is present, and gaining flow in a stream indicates that the unconsolidated sediments form an aquifer, for example along Shabbecong Creek in the vicinity of production well PVMSW01. In general, water-level and drilling-log data indicate that, except for the area close to Pohatcong Creek, only the bottom 10 to 20 ft of the unconsolidated sediments are saturated. Analysis of water levels in the surficial aquifer (the sediments), the carbonate-rock aquifer, and nearby streams indicate that the surficial aquifer is primarily a conduit for downward vertical flow of recharge from direct precipitation, runoff from the valley walls, and losing streams that originate in the crystalline-rock part of the valley. Close to Pohatcong Creek the surficial unconsolidated sediments typically have sufficient saturated thickness to function as an aquifer with a horizontal component of flow towards the creek as well as upward vertical flow created by discharge from the carbonate-rock aquifer.

The crystalline rocks bounding the northwestern and southeastern sides of the Pohatcong valley are virtually a no-flow boundary for ground water. Data supporting this conclusion include low specific capacities in wells completed in crystalline rock, and discharge to the small stream that drains the entire length of an approximately 3,000-ft-long tunnel at the head of the valley. Flow in the outlet stream was only about 0.5 ft^3/s on August 10, 2000. The small volume of ground water entering the tunnel and leaving by the outlet stream indicates there is little recharge to, and ground-water flow in, the crystalline rocks overlying the tunnel. The median specific capacity of 35 wells open to Precambrian crystalline rocks is 0.43 gal/min/ft, about one-quarter of the median specific capacity of wells open to the carbonate formations (1.7 gal/min/ft for 49 wells). A small amount of ground water may flow directly into the carbonate-rock aquifer from the crystalline rocks, but water-producing fractures typically decrease with depth, occurring infrequently at depths greater than about 500 feet below land surface. Because the surface of the crystalline formations is as much as 500 feet above the surface of the adjacent carbonate formations, recharge to the crystalline rocks is likely to follow shallow flow paths and discharge to upland streams and springs.

The thrust faults separating the crystalline and carbonate rocks could be a barrier to ground-water flow. Faults are high-permeability zones if rocks in the fault zone are competent but more fractured than the surrounding rock. Faults can act as a barrier to flow, however, if the rocks in the fault zone are more weathered than the surrounding rock, resulting in clays infilling fractures and creating a zone of low permeability. Well PVWPC13 (in a location shown in fig. 1 as overlying the Allentown Dolomite in fig. 2) probably penetrates rocks near or in the Pohatcong thrust fault. Although most wells in the crystalline rock penetrate 10 to 40 ft of unconsolidated sediments, this well penetrates 300 ft of saprolite, a significant barrier to flow. Similarly, a deep well on the edge of the valley near High Point Landfill (fig. 1) penetrates 327 ft of unconsolidated material, primarily clay, which may indicate it penetrates the Pohatcong fault. However, drawdown from nearby pumping in the crystalline rocks did affect water levels in a well (PVTVN13) that may have partly penetrated the Pohatcong Fault; consolidated rock was encountered over the entire length of this borehole. Water levels in wells open to the crystalline rocks on both the northwest and southeast sides of the valley typically are much higher than those in nearby wells open to the carbonate formations, indicating the two aquifers are separated by low permeability material. Whether this low permeability material is the thrust fault or the crystalline rock, the result is the same.

Overland runoff and base flow from the crystalline rocks probably increase ground-water recharge to the surficial aquifer (and hence the carbonate-rock aquifer) along the valley walls. Increased recharge along the valley walls is a significant feature in the same geologic setting of Long Valley to the southeast of the Pohatcong valley (Nicholson and others, 1996). In the Pohatcong valley, water emanating from springs has been observed along the valley walls; the water re-infiltrates into the unconsolidated sediments overlying the carbonate rock. Base flow in Brass Castle Creek decreased 0.10 ft^3/s in July 2000 and 0.68 ft^3/s in August 2000. This small loss of flow indicates that recharge to the surficial aquifer, and probably the underlying carbonate-rock aquifer, occurs in Brass Castle Creek. However, measurements made at two smaller, shorter tributaries to Pohatcong Creek demonstrated that the tributaries were not losing flow, indicating that recharge from losing streams is less significant in this valley than some others in northern New Jersey because of the low vertical hydraulic conductivity of the surficial aquifer.

Model-Grid Design

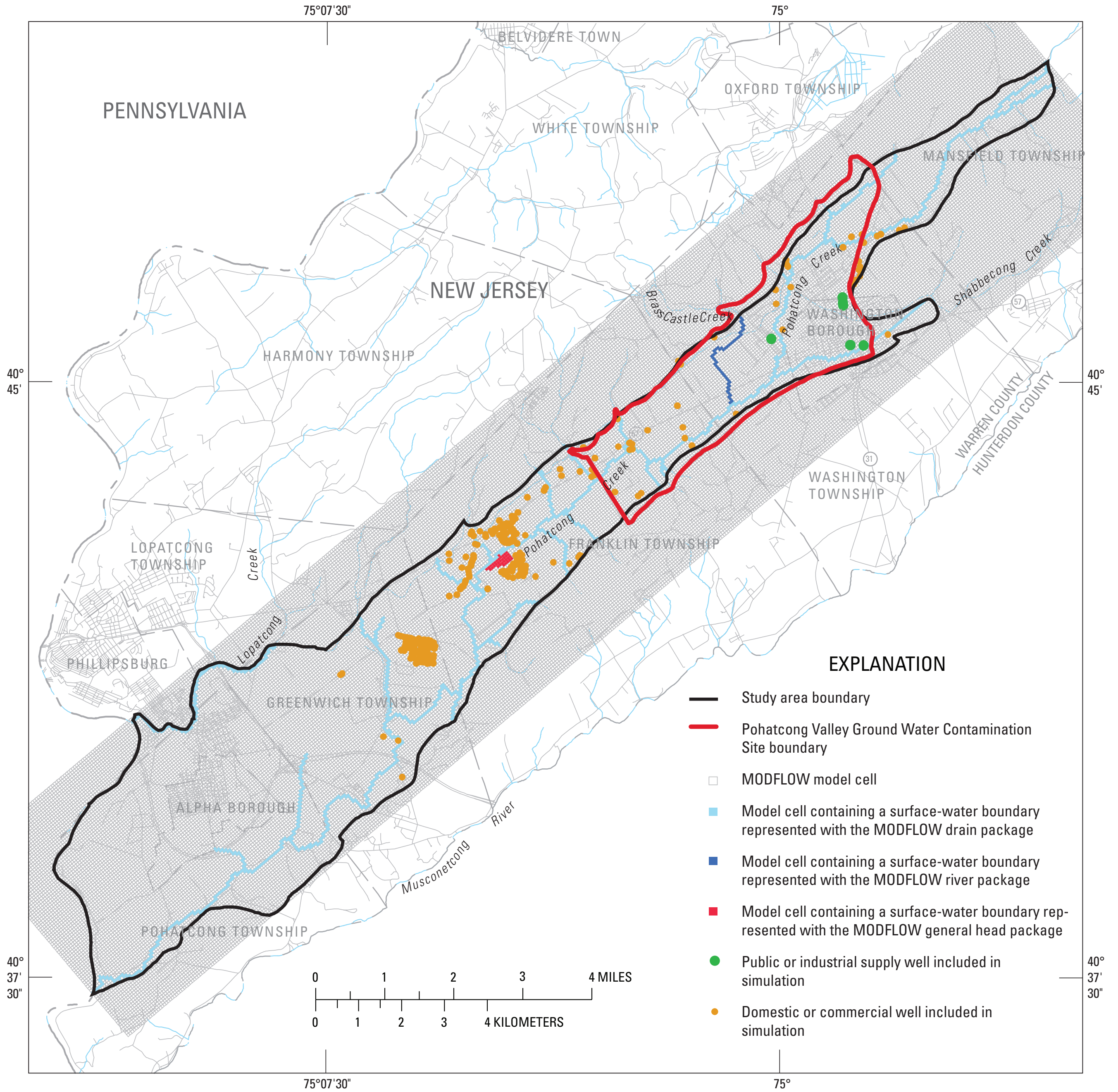
Ground-water flow in the carbonate-rock and surficial aquifers was simulated using the MODFLOW-2000 code (Harbaugh and others, 2000 and Hill and others, 2000). The model is a three-dimensional, finite-difference representation of ground-water flow in the Pohatcong Valley. Pre- and post-processing of the data were done with the Argus ONE (Argus Interware Inc., 1997) graphical user interface operating the USGS plug-in extension (Winston, 2000) and with the ESRI ArcMap geographic information system (GIS) software.

The model grid (fig. 14) has 92 rows and 508 columns (46,736 cells). The grid is uniform, with cells about 200 ft (61 m) on a side. A finer model grid was constructed with 100-ft cells (185 rows, 1,016 columns, 187,960 cells), but its use did not significantly change results of simulations. The coarser grid was chosen to make visualization of model input and output more readable. However, the finer grid spacing was used for the solute-transport simulation and could also be used in other simulations requiring greater detail. The grid is aligned parallel to the axis of the valley, at an angle of 50° east of north. The direction of maximum transmissivity of the bedrock is aligned approximately along the axis of the valley; therefore the anisotropy of the bedrock is represented in the model with maximum transmissivity along rows and minimum transmissivity along columns. The model has four layers (fig. 15). The top layer represents the surficial aquifer, and the bottom three layers represent the carbonate-rock aquifer. The surficial aquifer is modeled as an unconfined aquifer (with rewetting); therefore, many cells away from Pohatcong Creek are not active.

The carbonate-rock aquifer is modeled with three layers as confined and of constant thickness (total thickness about 500 ft (150 m), with each layer about 165-ft thick), resulting in constant transmissivity within each hydraulic conductivity zone. The aquifer was represented with three model layers because most of the aquifer-property tests (Carleton and others, 2005) were done in the upper 160 feet of the aquifer, and calibration results (described farther on in this report) were improved by representing the carbonate-rock aquifer with three layers instead of one. Specifically, simulated water levels for most of the valley more closely matched measured water levels when model layer 2 (fig. 15) had transmissivity values similar to those measured in the upper 160 ft and model layers 3 and 4 had lower transmissivities. The aquifer is presumed to have decreasing permeability with depth and negligible permeability below a depth of 500 ft (see for example Nicholson and others, 1996; Lewis-Brown and Jacobsen, 1998; and Carleton and others, 1999). The exact function that describes decreasing permeability with depth is not known. For this study, it was assumed that the horizontal and vertical hydraulic conductivities of model layer 3 are three times those of layer 4 and one-third of those of layer 2.

Boundary Conditions

Virtually all of the boundaries of the model are natural hydrologic boundaries. The lateral and bottom boundaries are no-flow boundaries (fig. 15). The upper boundary of the model includes recharge (applied to the topmost active cells) and the Delaware River, streams, and one abandoned, water-filled quarry (simulated with the river, drain, and general head boundary packages of MODFLOW, respectively).



Base from U.S. Geological Survey digital line graph files, 1:24,000

Figure 14. Extent of grid and boundaries used in the numerical simulation of ground-water flow, Pohatcong Valley, Warren County, N.J.

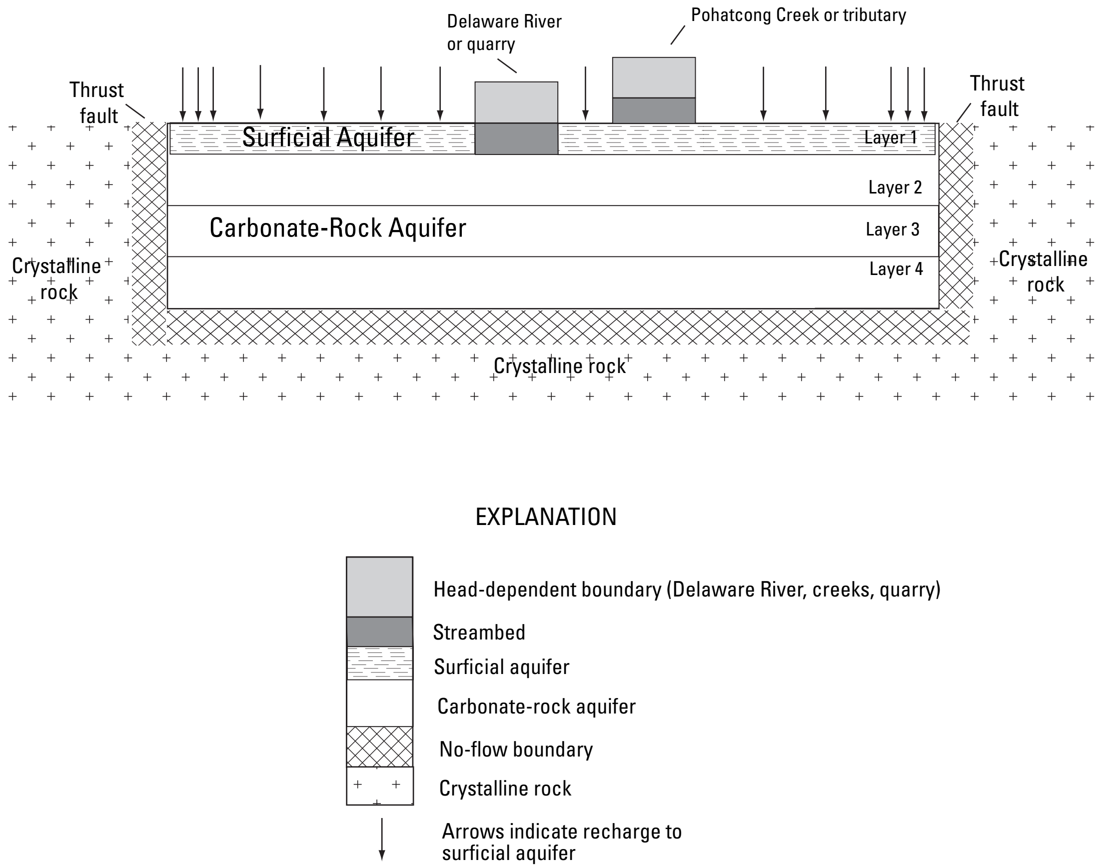


Figure 15. Schematic representation of simulated flow system with model layers, Pohatcong Valley, Warren County, N.J.

Lateral Model Boundaries

The lateral boundaries (excluding the southwestern end of the model at the Delaware River) are the contacts between the carbonate-rock aquifer and the Precambrian crystalline rock or, in the Shabbecong Valley, the Cambrian Hardyston quartzite. Although some ground water undoubtedly flows across this boundary, the flux is considered to be negligible compared to flow within the carbonate-rock aquifer (see section “Conceptual model of ground-water flow”). All of the surface-water boundaries represented in the model (except the quarry), including Pohatcong, Shabbecong, and Brass Castle Creeks and the smaller tributaries, originate in Precambrian crystalline-rock parts of the Pohatcong Valley. All precipitation falling in the crystalline-rock areas of the valley is assumed to discharge to surface water (either as runoff or via ground-water discharge) outside the domain of the model and, therefore, is not explicitly modeled, with the exception of additional recharge added to the model along valley walls to account for runoff from crystalline-rock areas entering the carbonate-rock aquifer (described in “Recharge” farther on).

The lateral boundary at the southwestern end of the valley is the Delaware River. Specifically, the lateral boundary is no-flow, and the upper boundary along the border is the Delaware River (modeled with the general head boundary package in MODFLOW), but the result is virtually the same as having a lateral boundary of constant head cells. Although the river does not fully penetrate the carbonate-rock aquifer and theoretically, underflow is possible, it is believed that the river acts as a regional sink, and no ground water from the Pohatcong Valley flows beneath the river.

A small section of the lateral boundary along the northwestern valley wall is not a simple hydraulic boundary. From just southwest of New Village to the Delaware River, the carbonate rocks are continuous from the Pohatcong Valley into the next valley to the north, the Lopatcong Creek Valley. Where the contact between the crystalline and carbonate rocks turns to the north (near New Village), the model boundary approximately follows the land contour, drops a short distance to Lopatcong Creek, then follows Lopatcong Creek to the Delaware River. Because of the anisotropy of the carbonate rocks, it is unlikely that a significant amount of ground water flows northwest from the Pohatcong Valley towards the Lopatcong Valley (flowing instead downvalley to the Delaware River) and very unlikely that any flow towards the northwest continues under Lopatcong Creek. Therefore, this lateral boundary is considered to be representative of the system for the purposes of these simulations.

Surface Water

Lopatcong Creek and Pohatcong Creek and most of its tributaries are modeled with the drain and river packages of MODFLOW. The river package allows water to exit or enter the model domain, depending on whether the head in a cell including a river is greater than or less than, respectively, the designated river stage. The drain package allows water to leave the model domain if head in a cell containing a drain is higher than the designated drain stage. The stages of river segments (fig. 12) were determined for points where cross sections were marked on Federal Emergency Management Agency (FEMA) flood maps for Greenwich, Franklin, Washington, and Mansfield Townships and Washington Borough (Federal Emergency Management Agency, 1982a, 1982b, 1982c, 1982d, 1983). Where no flood-map data were available (which occurred only for tributaries), stages were estimated from 20-ft land-surface-altitude contours on USGS 7.5-minute quadrangle maps. River segments were assigned a stage from the midpoint between individual point estimates of stage. The altitude of the bottom of the river (the streambed) is needed for determining whether flux from the river varies with head changes in the aquifer: if the head in the aquifer falls below the bottom of the river, the flux is constant regardless of aquifer head, a function only of the specified stage of the river and bed conductance. The altitude of the bottom of the river segments was assumed to be 3 ft below the river stage. During initial calibration of the model, all streams were simulated with the MODFLOW river package, but the discrete river stages derived from the FEMA flood maps caused large flows into or out of the model domain in cells representing areas just upstream or downstream from a stream stage change, respectively. Because all streams in the study area (except Brass Castle Creek) are believed to be gaining, or at least not losing, it was appropriate to remove the artifact of discretization-related river fluxes by using the drain package. Because Brass Castle Creek had measured losses, it was represented with the river package.

The flux between the aquifer and the river is a function of the head gradient and the conductance of the riverbed. Bed conductance is defined as the area (width times length of the river reach in each model cell) times hydraulic conductivity of the bed material divided by the thickness of the bed. The length of each river segment (if any) in each model cell was determined by the geographic information system (GIS) functions of the pre-processing software. The widths of Pohatcong Creek and its tributaries were approximated on the basis of widths measured at the time of the base-flow measurements in 2000. Pohatcong Creek was assumed to be 10 ft wide from its headwaters down to station 01455135 (fig. 8), 15 ft wide from station 01455135 to the confluence with Shabbecong Creek (near Rt. 57), 20 ft wide from Shabbecong Creek to just upstream from Broadway, and 30 ft wide from just upstream from Broadway to the confluence with the Delaware River (at Carpentersville). Brass Castle Creek, Shabbecong Creek, and the remaining tributaries are assumed to be 10, 6.5, and 3 ft wide, respectively. Lopatcong Creek is assumed to be 20 ft wide. The thickness of the riverbed was assumed to be 1 ft. The hydraulic conductivity of the riverbed was not measured but was initially estimated to be about 5 ft/d. The thickness and hydraulic conductivity of the riverbed were not known, and stream widths were approximated between locations where measurements were made; therefore, bed conductance is considered to be a lumped parameter. The lumped parameter of bed conductance was varied for measured stream reaches during model calibration to achieve optimal results. When the stream widths and bed thickness described above are factored out of the lumped parameter, then the calibrated riverbed hydraulic conductivities used for Pohatcong Creek is 26 ft/d from its headwaters downstream to station 01455135, 3.7 ft/d from station 01455135 to the confluence with Shabbecong Creek (near Rt. 57), 1.9 ft/d from Shabbecong Creek to the confluence with Brass Castle Creek, 0.93 ft/d from Brass Castle Creek down to just upstream from Broadway, and 0.13 ft/d from just upstream from Broadway to the confluence with the Delaware River (at Carpentersville). The decreasing hydraulic conductivity in the downstream direction might be attributable to the increase of fine sediments in the downstream direction (which would be expected as the stream widens and slows). The calibrated riverbed hydraulic conductivities of Brass Castle Creek, Shabbecong Creek, and the remaining tributaries are 0.4, 1.2, and 1.2 ft/d, respectively.

The Delaware River was modeled using the general head boundary package of MODFLOW. The general head boundary package is the same as the river package, except there is no provision for keeping flux constant if the head in the aquifer falls below the bottom of the streambed. Use of the general head boundary package simplified the identification of simulated ground-water discharge to the Delaware River rather than to the smaller streams. As in the river package described above, the conductance term is a lumped parameter containing the variables of length, width, riverbed thickness, and river-bed hydraulic conductivity. The width and bed thickness were assumed to be 500 ft and 5 ft, respectively. The river was divided into 20 segments with stage ranging from 140 to 152 ft (42.8 to 46.5 m), in increments of about 0.65 ft (0.2 m). The stage of the river input to the model was interpolated from the 20-ft contours on the Easton USGS 7.5-minute quadrangle map. The best-fit riverbed hydraulic conductivity was 0.001 ft/d, a lower value than other riverbed hydraulic conductivities used, but not unreasonable given the low sensitivity of the model to this parameter, lack of flow data, and modeling the Delaware River as being directly connected to the bedrock (to avoid problems that occurred when layer 1 was modeled as unconfined and cells were prone to become inactive) rather than the low permeability surficial-aquifer sediments.

Ground-Water Withdrawal and Injection Wells

Production, industrial, commercial, and domestic production wells were modeled with the MODFLOW well package. These wells remove ground water from the model domain except for injection well P485, which introduces water to the model domain. Withdrawal or injection rates are listed in table 6. The six wells listed in table 6 that are completed in the carbonate-rock aquifer are the production wells PVMSW01 and PVMSW04, commercial well (C493), and industrial wells P483, P484, and P485. Hundreds of domestic wells outside the well-exclusion zone established by the NJDEP, primarily southwest of Broadway, withdraw water from the carbonate-rock aquifer. The location and withdrawal rate of every domestic well is not known. Domestic wells identified by the USGS, CH2M Hill, or the NJDEP for this study, primarily located from Broadway to Stewartsville (fig. 1), were included in the model. Withdrawal rates for domestic wells were estimated by determining approximate usage by production customers. The average annual withdrawals from the three production wells serving the Pohatcong Valley divided by the number of commercial and domestic customer hook-ups served by those wells (Frank Hadley, New Jersey American Water Company, oral commun., 2006) yields an estimate of average use of 206 gallons per day per customer hook-up. However, most properties with self-supply wells discharge wastewater on site; therefore, some of the water withdrawn from the carbonate-rock aquifer by domestic wells is returned to the surficial aquifer by way of the septic system. The percentage of domestic well water ultimately returned to the ground-water system by septic systems is not known, but estimates range from 20 to 80 percent. For this study, it was assumed that 40 percent of the water was consumed (about 85 gallons per day per well). To estimate the potential effect on the ground-water system of domestic withdrawals, an order-of-magnitude comparison of the withdrawals from domestic wells was made with those from production wells: withdrawals from the two production wells in the Pohatcong Valley equal withdrawals from about 2,330 homes or, if 60 percent of withdrawn water is returned to the ground-water system, about 5,640 homes with septic systems. The effect on flow paths from such diffuse withdrawals as those in New Village (less than 300 homes) and farther down the valley is negligible.

Recharge

Ground-water recharge was nonuniformly distributed over the model area to the topmost active layer. Most cells were assigned a uniform recharge from precipitation, but some cells on the perimeter of the model received additional recharge representing runoff from upland subbasins (areas within the Pohatcong Creek basin underlain by crystalline rock and, therefore, not included in the model). If a subbasin contains no upland tributary, runoff from this area was applied to the valley-perimeter cells as additional recharge. If the upland area of a subbasin drains directly to a Pohatcong Creek tributary outside the modeled area, that water was available to the model domain from leakage from the tributary. Therefore, only a small amount of direct runoff from the upland area was assumed available to infiltrate at the carbonate-rock/crystalline-rock boundary, and a small increase over minimum recharge was applied to these perimeter cells. The recharge to the model from the upland areas was determined by (1) dividing the study area into surface-water subbasins , (2) determining the upland area that contributes runoff to the valley in each subbasin, (3) multiplying the contributing upland area by the average unit-area base flow and, (4) dividing this flow by the number of perimeter cells of the modeled area adjacent to the subbasin and applying the result to these cells (fig. 16). During calibration, the extra recharge from outside of the modeled area was reduced in the Shabbecong Valley area because simulated water levels were too high, perhaps because the ratio of linear feet of area receiving extra recharge to square feet of adjacent modeled area was large compared to the rest of the modeled area. The rate of recharge was uniformly increased or decreased during calibration and the final rate was 10 percent higher than initial values used.

The variation in the rate of recharge applied to the model is shown in figure 16. Estimates for the throughflow term (T) and the direct discharge to the Delaware River (B_D) in the recharge equation (described in the section "Water Budget") are not based on direct measurement but on assumptions that substantially affect the results; therefore, the recharge (R) was estimated during model calibration. Recharge in most model cells is a uniform 10 in/yr, with higher amounts (typically about 50 in/yr with a maximum of 130 in/yr) applied in model cells adjacent to the lateral boundary believed to receive additional recharge from upland runoff. The final value for basin-wide recharge is 45 ft^3/s (11 in/yr over the entire basin, 20 in/yr over the carbonate-rock part of the basin), the same as the estimated mean annual base flow for Pohatcong Creek at Carpentersville, and about 30 percent lower than the estimate made in the "Water Budget" section of this report.

Aquifer Hydraulic Conductivity

The ground-water-flow model was used to estimate hydraulic properties over the extent of the valley that were not well known, including the horizontal and vertical hydraulic conductivities of the surficial and carbonate-rock aquifers. No field tests were done to estimate vertical hydraulic conductivities or the horizontal hydraulic conductivity of the surficial aquifer (although specific capacity data indicate the median hydraulic conductivity is about 3 ft/d). The horizontal hydraulic conductivity and horizontal anisotropy of the carbonate-rock aquifer were estimated from analysis of data from aquifer tests and packer tests, but these estimates ranged widely and most packer tests were conducted less than 100 feet into the bedrock. The horizontal anisotropy of the carbonate-rock aquifer was calculated (from a large-scale aquifer test) to be 0.46 and calculated horizontal hydraulic conductivities from many aquifer tests range from 0.05 to 1,800 ft/d. Although the lower horizontal hydraulic conductivity values might represent most of the volume of the aquifer, any interconnected zones of higher hydraulic conductivity would overwhelmingly domi-

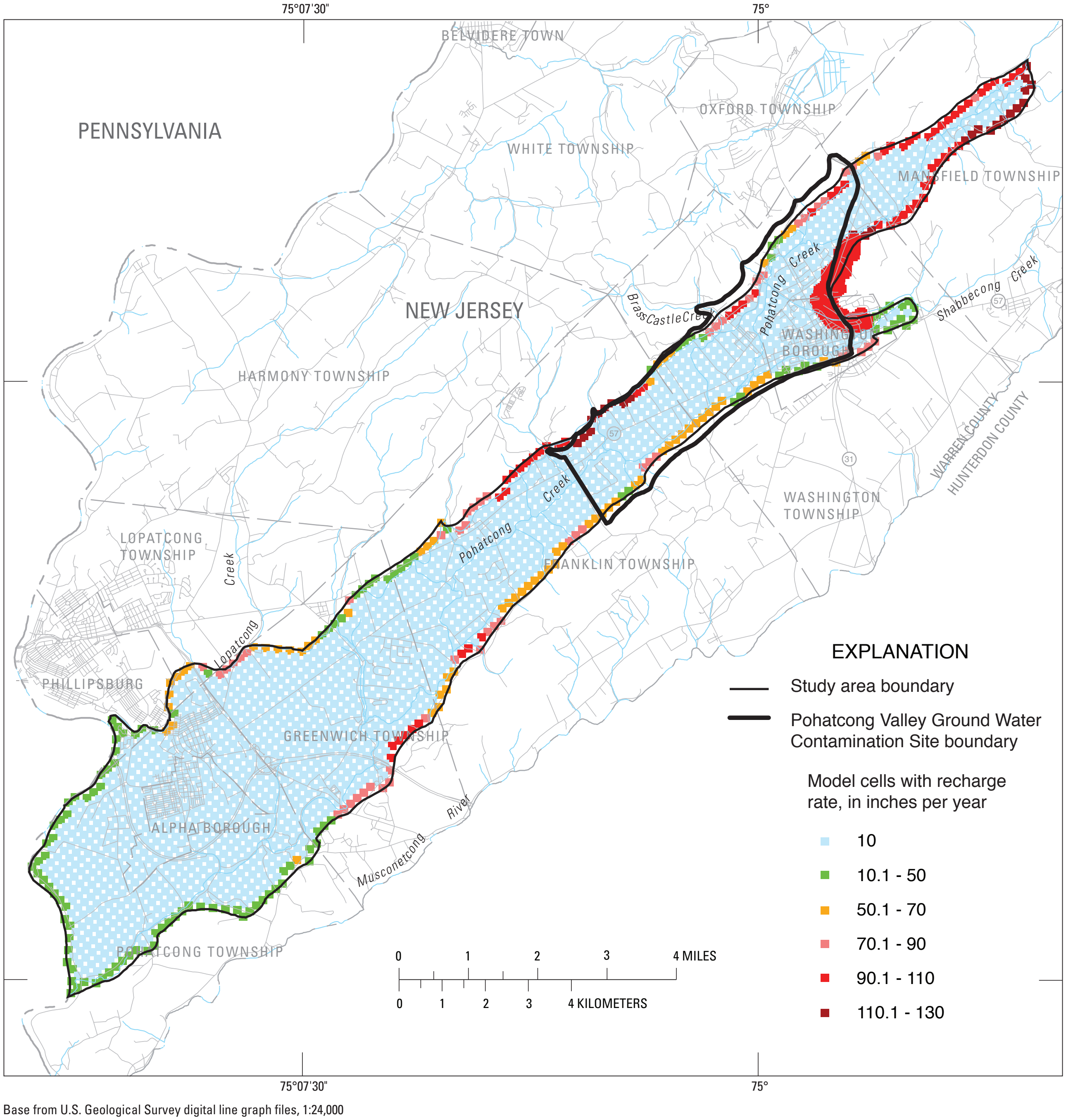


Figure 16. Distribution and magnitude of recharge applied to the numerical ground-water-flow model, Pohatcong Valley, Warren County, N.J.

nate the system; therefore, the uniform values applied in the model were set towards the higher end of the range.

The horizontal and vertical hydraulic conductivities were not varied over the model domain except where model calibration was improved by assigning different values to areas with differing hydrogeology. Water levels close to the crystalline-rock boundary were high relative to water levels elsewhere in the valley, possibly the result of low horizontal hydraulic conductivity caused by weathering of the boundary faults. Therefore, a lower horizontal hydraulic conductivity was assigned to a thin zone around the perimeter of the carbonate-rock aquifer. Simulated water levels were higher than measured levels in the central valley from the vicinity of PVCHU02 to southwest of Broadway, and lower than measured near New Village and in the upper Pohatcong Valley (in the Shabbecong valley, near P483 and P484 , and northeast of PVCHU02). Decreasing hydraulic conductivity generally will increase water levels; therefore, the hydraulic conductivity assigned to the area in the upper Pohatcong Valley underlain by the Leithsville Formation is lower than that assigned to the area underlain by the Allentown Dolomite (fig. 17). This is supported in part by the results of the aquifer tests using PVMSW01 (Leithsville Formation) and PVMSW04 (Allentown Dolomite), which yielded horizontal hydraulic conductivities of 34 ft/d and 61 to 234 ft/d, respectively. Similarly, hydraulic conductivities slightly (30 percent) lower than those in the main valley were assigned to an area roughly corresponding to the area underlain by the Beekmantown Group and Jacksonburg Limestone near and downvalley from New Village. On the basis of water-level data, this lower hydraulic conductivity zone was not extended as far upvalley as the formations are shown to extend on the geologic maps. The discrepancy was considered acceptable because contacts shown on the geologic maps are generally approximate (because the carbonate rock is buried throughout most of the valley).

The final simulated horizontal and vertical hydraulic conductivities for the surficial aquifer are both 6.6 ft/d. Typically unconsolidated sediments are assumed to have a horizontal to vertical hydraulic conductivity ratio of 10 to 1, but a slightly better calibration was achieved using a 1 to 1 ratio. The model was insensitive to horizontal hydraulic conductivity and slightly sensitive to vertical hydraulic conductivity. It was not designed to closely simulate the surficial aquifer; therefore, the 1 to 1 ratio is considered acceptable. The final maximum hydraulic conductivities for the three layers representing the carbonate-rock aquifer are 300, 100, and 33 ft/d (horizontal) and 100, 33, and 11 ft/d (vertical) for layers 2, 3, and 4, respectively. The 3 to 1 ratio for horizontal to vertical hydraulic conductivity of the carbonate-rock aquifer is considered appropriate because the steeply dipping bedding planes of the formations are believed to promote greater vertical connectivity than would be the case with flat-lying bedding planes. Carbonate-rock horizontal and vertical hydraulic conductivities in the area corresponding to the Leithsville formation were multiplied by 0.08, in areas corresponding to the Beekmantown Group and Jacksonburg Limestone by 0.7, and around the perimeter by 0.018 (fig 17). The final horizontal anisotropy assigned to the three carbonate-rock model layers was 0.2.

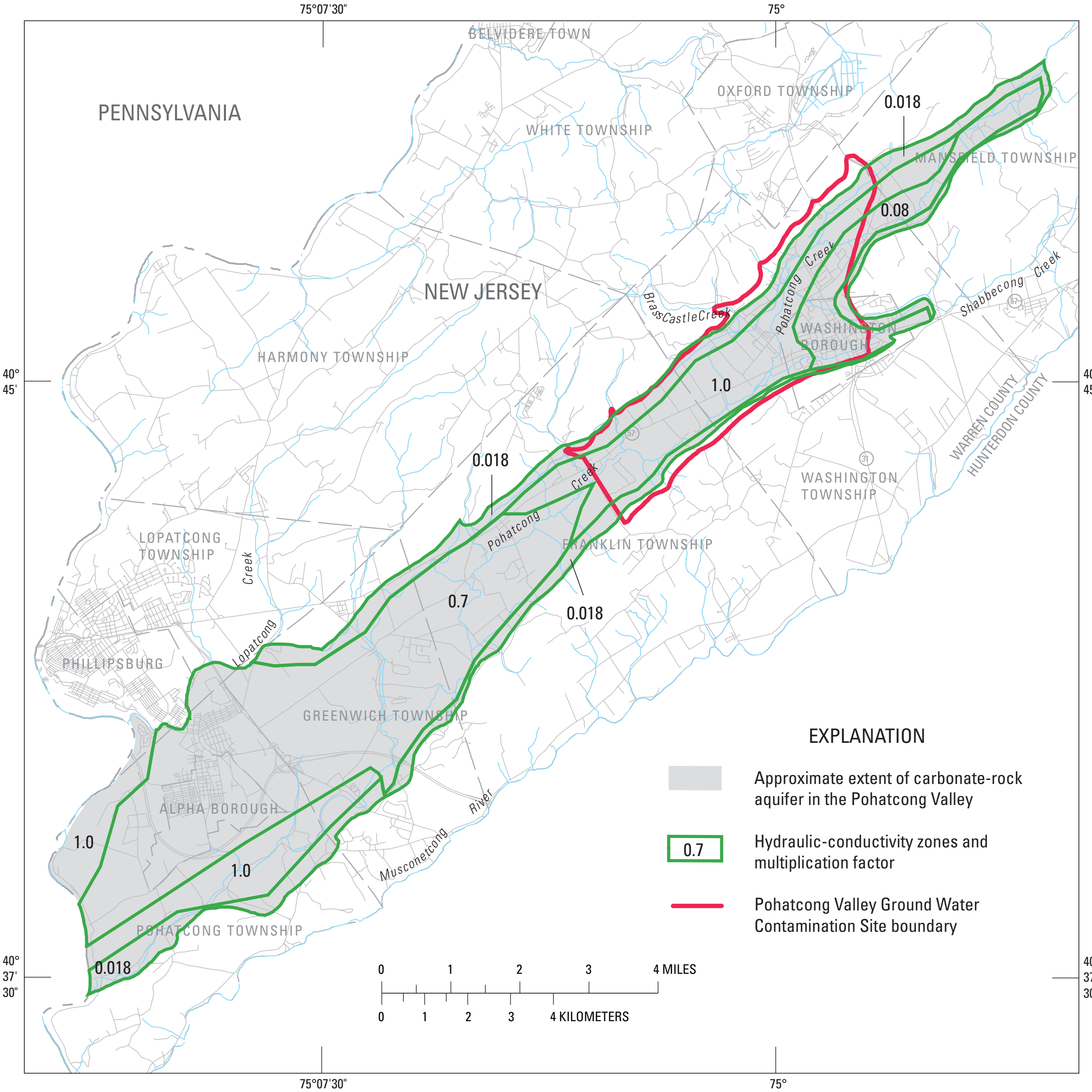
Model Calibration

The model was calibrated using a combination of automated parameter estimation and trial-and-error techniques. Simulated and measured water levels, discharge to streams, and ground-water travel times were compared in order to determine the combination that resulted in the lowest cumulative error. The model was most sensitive to changes in recharge and horizontal hydraulic conductivity of model layer 2; somewhat sensitive to horizontal anisotropy of layer 2, horizontal hydraulic conductivity of layer 3, and vertical hydraulic conductivity of layer 1; less sensitive to bed conductance of Pohatcong Creek, horizontal anisotropy of layer 3, and horizontal hydraulic conductivities of layers 4 and 1; and insensitive to horizontal anisotropy of layer 4, vertical hydraulic conductivity of layers 2, 3, and 4, bed conductance of the Delaware River, Brass Castle Creek (simulated with the river package), other Pohatcong Creek tributaries (simulated with the drain package), and the quarry (fig 18).

During automated calibration, weights were assigned to each water-level or base flow observation according to a strict formula (Hill, 1998), but this approach assigned less importance to base-flow data than was considered appropriate because (1) there were far fewer base-flow measurements than water-level measurements and (2) the inherent errors in calculating gain or loss for a stream reach reduce the calculated weight compared to precise water-level measurements made in surveyed wells (for which the uncertainty of error with respect to mean annual water level was not known). Therefore, during manual calibration, selected base-flow and water-level residuals were informally assigned greater weight. For example, water levels near the TCE source area were considered important and assigned greater weight, leading to poorer matches with water levels measured in wells in the Shabbecong Valley. Similarly, base flow to Pohatcong Creek in the reaches from the vicinity of the TCE source area to Broadway were more heavily weighted during calibration because these results could substantially affect where flow paths originating in the TCE source area terminated along Pohatcong Creek.

Simulated Water Levels

Simulated water levels were compared to water levels measured in 56 wells open to the carbonate-rock aquifer. Water-level measurements were made in 137 wells in the Pohatcong Valley and about 100 of the altitudes were determined to be accurate to within 0.3 ft. Some wells were not used for model calibration because they were open to the crystalline-rock or surficial aquifers or were considered redundant because they were close to another well. During automated parameter estimation, water levels were assigned a



Base from U.S. Geological Survey digital line graph files, 1:24,000

Figure 17. Zones of horizontal and vertical hydraulic conductivity in the carbonate-rock aquifer used in the numerical simulation of ground-water flow, Pohatcong Valley, Warren County, N.J.

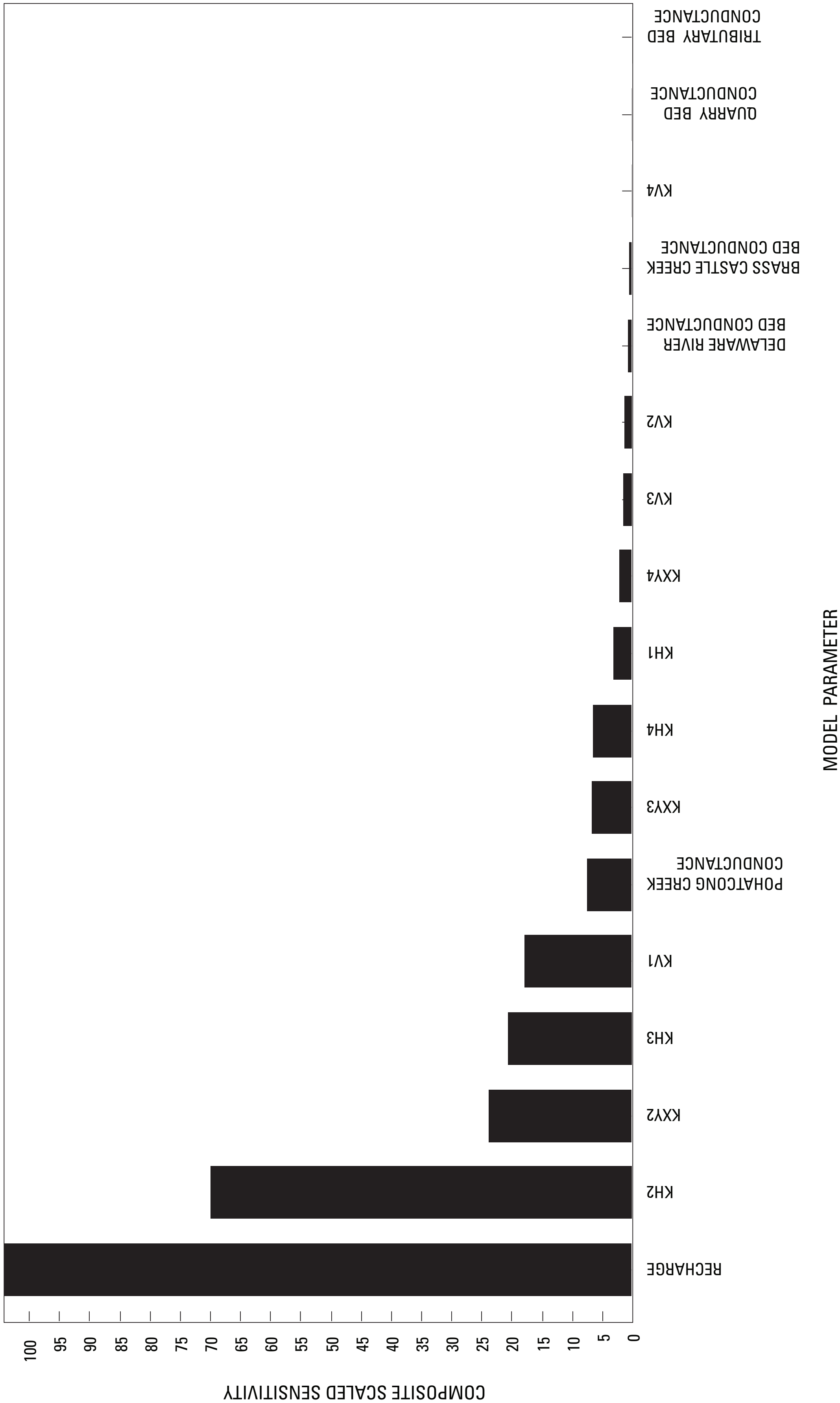


Figure 18. Composite scaled sensitivity of ground-water-flow model output to changes in input parameters, Pohatcong Valley, Warren County, N.J. (KH, horizontal hydraulic conductivity; KV, vertical hydraulic conductivity; KXY, horizontal anisotropy ratio of hydraulic conductivity. Numbers are model layer.)

weight according to the standard deviation of the measurement error (Hill, 1998, p. 46)-- 0.16 ft for surveyed monitoring or domestic wells, 0.59 ft for well C487 (altitude estimated to be within 1 ft), and 0.55 ft for wells in which pumps were operating at the time the water levels were measured (P485 and PVMSW04). Water levels from the June 2001 synoptic study were used because those data were considered to be more representative of average annual conditions than data collected in June 2002, near the end of the 2001-02 drought. Representative water levels in the surficial aquifer were difficult to determine because of significant differences between the deep and shallow parts of the aquifer. Therefore, water levels in the surficial aquifer were not included in the calibration process. Water-level residuals (measured – simulated water-level altitudes) for the carbonate-rock aquifer are shown in figure 19 and table 7, and simulated water-level contours are shown in figure 20.

Statistics were calculated for the water-level residuals, including median, -3.28 ft; mean, -2.73 ft; root mean square error, 1.13 ft; maximum, 14.99 ft; and minimum, -23.56 ft. Thirty-four water-level residuals are less than zero, 22 are greater than zero, and 50 percent range from -6 ft to 6 ft. Simulated water levels generally were lower than the measured levels in the vicinity of the quarry, low along the valley walls, high from northeast of New Village to the vicinity of the Washington Borough border in the center of the valley, high in the Shabbecong Valley, and low near the TCE source area. A cross-section of measured water levels in the vicinity of Broadway has a pronounced U-shape, with water levels near the valley wall as much as 40 ft higher than in the center of the valley. Simulated water levels do not have as pronounced a U-shape and generally are too low near the valley wall and too high in the center of the valley. A more pronounced U-shape could be generated with higher horizontal anisotropy in the carbonate-rock aquifer, but anisotropies substantially higher than considered reasonable (based on analysis of aquifer tests) produced a modest improvement in cross-section shape at the expense of water levels elsewhere (for example in the Shabbecong Valley) and stream base flows. Addition of the narrow, low hydraulic-conductivity zone along the valley walls improved the calibration of the cross-section shape without degrading the calibration elsewhere. Use of higher horizontal conductivities for layers 2, 3, and 4 (carbonate rock) lowered simulated water levels and improved the fit to the water-level data, especially in the center of the valley, but conductivity values higher than those used were not considered feasible. Furthermore, regional lowering of simulated water levels increased residuals in the vicinity of the TCE source area, which was not considered an acceptable trade off.

Simulated water-level contours (fig. 20) are similar in shape and location to contours drawn from measured data (fig. 10). The long downvalley extent of simulated contours along the valley wall is caused by the zone of low hydraulic conductivity along the border of the model. The drawn contours do not share the exact shape, but are based on sparse data and might not be accurate along the valley walls.

Simulated Stream Base Flow

Stream base flow was measured in July and August 2000. The August measurement was considered to be closer to mean annual flow than the July measurement and, therefore, was used for model calibration. The gain or loss of flow in reaches of Pohatcong Creek and its tributaries was determined by subtracting, from the measured flow at a downstream station, the flow at the next upstream station and any throughflow from tributaries entering in that reach. Measurements were made at six locations on Pohatcong Creek; therefore, gain or loss could be calculated for five reaches. Because small tributaries were considered to be neither gaining nor losing, measured tributary flows were subtracted from Pohatcong Creek flows, and all interaction between the carbonate-rock aquifer and surface water was assumed to occur along the mainstem of Pohatcong Creek, Shabbecong Creek, and Brass Castle Creek. All tributaries upstream from New Village with discernable flow were measured in at least one location. Gains or losses for the reaches of Brass Castle and Shabbecong Creeks from the carbonate/crystalline rock boundary to the confluence with Pohatcong Creek were calculated. No measurements were made on Pohatcong Creek or its tributaries downstream from New Village; therefore, the gain of Pohatcong Creek from New Village to Carpentersville was calculated by subtracting the measured discharge at New Village from the estimated mean annual base flow at Carpentersville without considering tributaries. During the process of automated parameter estimation, less weight was assigned to base-flow calculations because of the compounded effect of measurement errors when subtracting one base-flow measurement from another. Base-flow observations were weighted by the coefficient of variation (Hill, 1998, p. 46-47) to allow streamflow gains and losses to be accurately weighted relative to water-level measurements. Base-flow residuals (measured – simulated) are shown in figure 19 and in table 7b.

Base flow on Pohatcong Creek at New Village, measured on August 8, 2000, was 84 percent of mean annual base flow estimated on the basis of low-flow correlations with nearby index stations. For consistency, the estimated mean annual base flow for Pohatcong Creek at Carpentersville was multiplied by 0.84 so that all calibration was to approximate or measured base-flow data of August 8, 2000. During calibration, however, simulated base flows slightly higher than measured flows were considered to be a better match to mean annual conditions than the measured base flows. Along Pohatcong Creek, simulated base flow was higher than estimated base flow at five sites and lower than measured at one site. Simulated base flow to Shabbecong Creek was higher than estimated base flow. Simulated loss of flow from Brass Castle Creek to ground water was less than the estimated loss.

No measurement of the discharge from the quarry was made. The estimated base flow resulted from two visual estimates of flow made by CH2M Hill personnel of 2 and 4 Mgal/d (3 and 6 ft^3/s) (Murray Rosenberg, CH2M Hill, oral commun., 2005). Therefore, simulated and estimated dis-

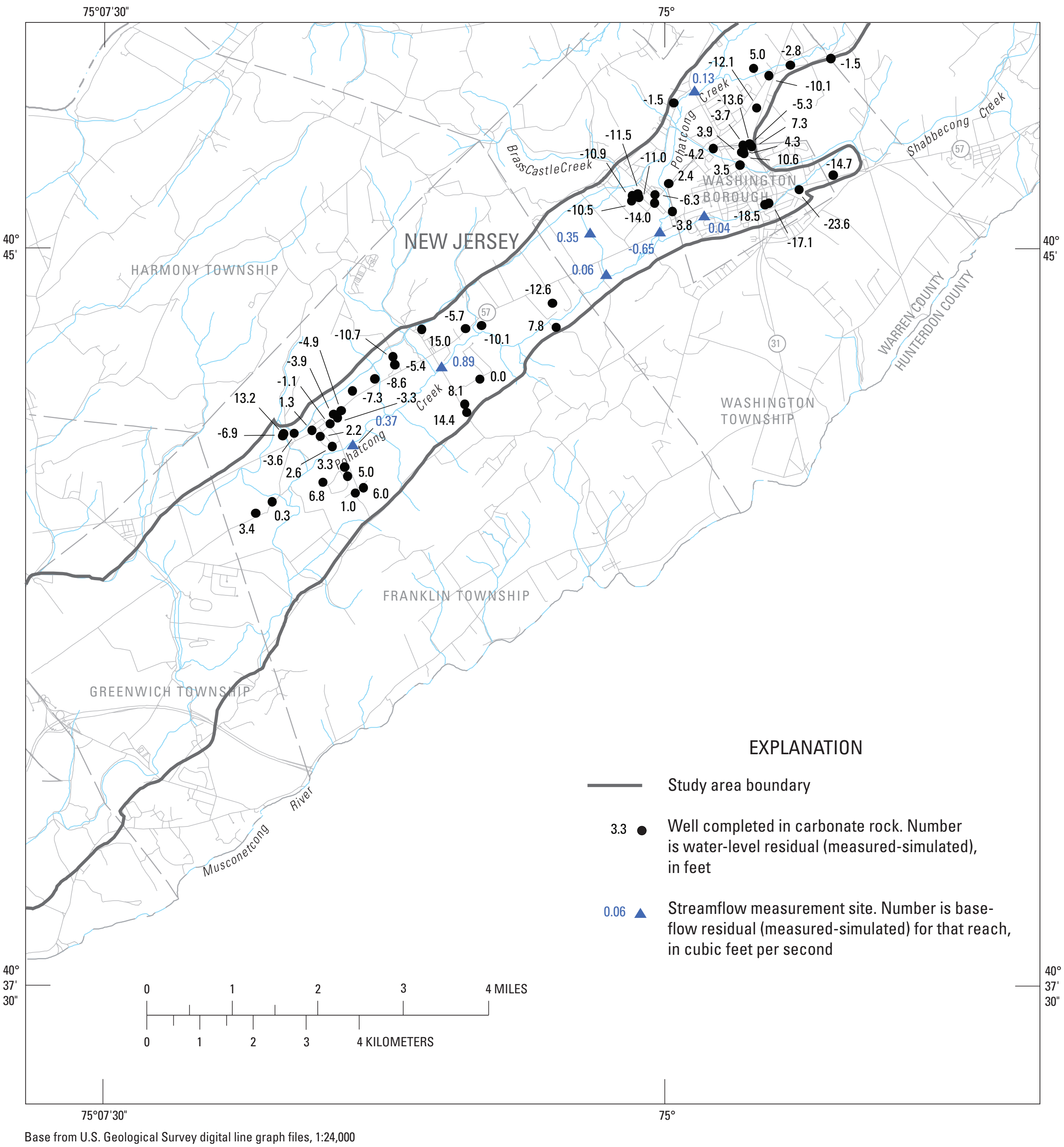
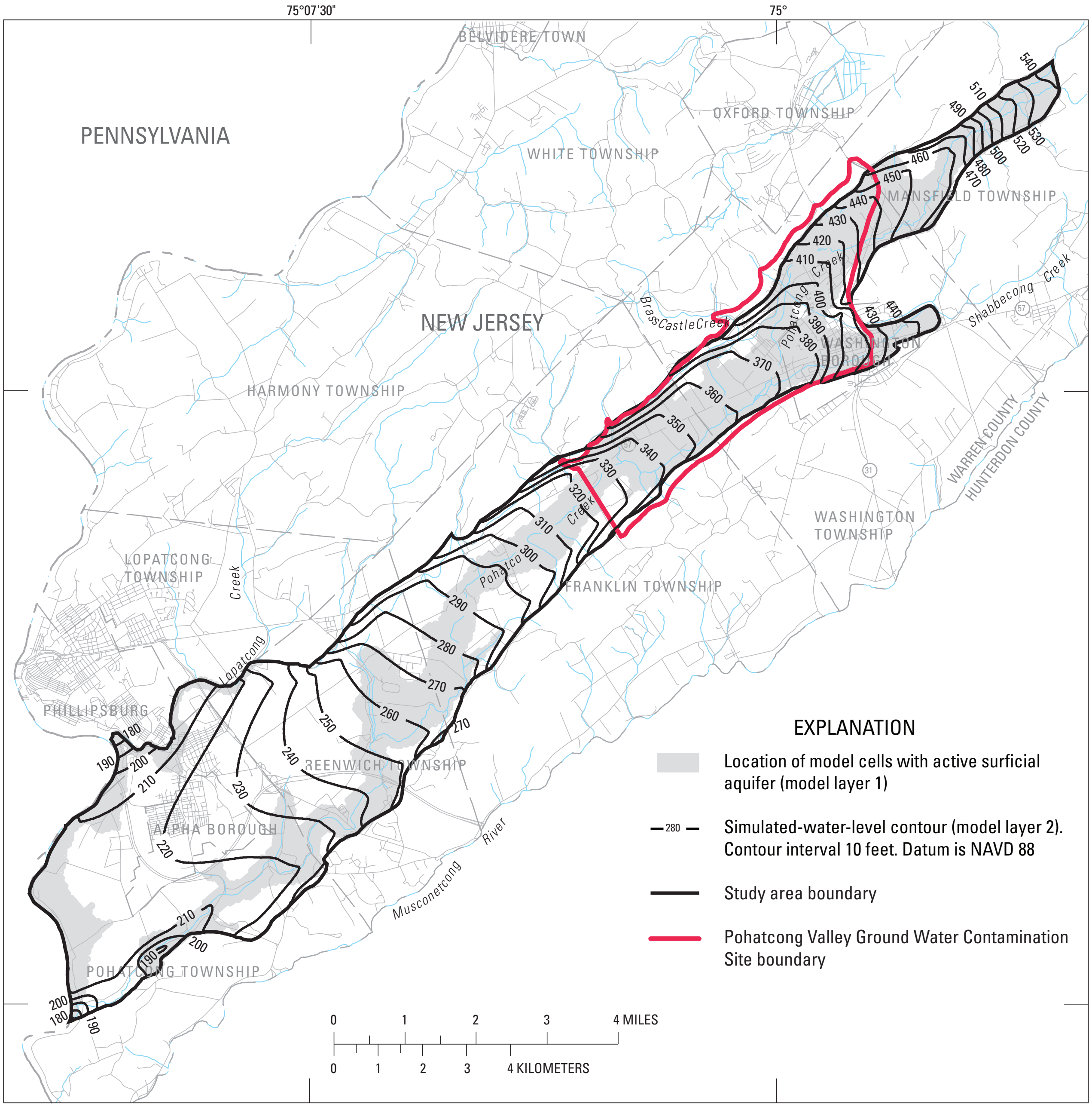


Figure 19. Water-level residuals (measured minus simulated) for wells completed in the carbonate-rock aquifer and stream base-flow residuals for Pohatcong Creek and selected tributaries, Pohatcong Valley, Warren County, N.J.



Base from U.S. Geological Survey digital line graph files, 1:24,000

Figure 20. Simulated water-level contours for the carbonate-rock aquifer and location of active surficial-aquifer model cells, Pohatcong Valley, Warren County, N.J.

Table 7a. Measured and simulated water levels in wells open to the carbonate-rock aquifer, Pohatcong Valley, Warren County, N.J.

[Water levels were measured June 2001; NAVD88, North American Vertical Datum of 1988]

Well identification code	Measured water level altitude (in feet, NAVD88)	Simulated water level altitude (in feet, NAVD88)	Residual (measured minus simulated) (in feet)	Well identification code	Measured water level altitude (in feet below NAVD88)	Simulated water level altitude (in feet below NAVD88)	Residual (measured - simulated) (in feet)
A3	438.67	433.72	4.95	D436	289.16	285.78	3.38
A470	369.18	354.19	14.99	D437	452.55	455.31	-2.76
A472	313.93	322.53	-8.60	MW-42D	372.14	383.66	-11.52
A474	389.31	386.87	2.44	MW-42E	371.29	381.82	-10.53
C487	416.06	439.62	-23.56	MW-42F	372.03	382.96	-10.93
C488	368.25	360.41	7.84	MW-42H	372.04	383.00	-10.96
D60	470.20	471.70	-1.50	MW453	339.11	339.10	0.01
D70	433.55	443.62	-10.07	MW454	349.13	341.06	8.07
D206	304.98	308.62	-3.64	P484	423.40	416.05	7.35
D211	304.68	301.41	3.27	P485	398.44	403.79	-5.35
D218	309.66	314.55	-4.89	PVANC03	425.79	439.44	-13.65
D225	307.02	304.45	2.57	PVANC05	427.59	431.26	-3.67
D234	308.48	306.25	2.23	PVANC06	425.53	421.66	3.87
D238	306.38	301.36	5.02	PVANC29	418.60	414.30	4.30
D242	304.82	311.68	-6.86	PVANC30	423.33	412.73	10.60
D244	303.43	296.61	6.82	PVCHU02	376.24	382.54	-6.30
D246	308.00	306.68	1.33	PVFCC12	377.90	381.67	-3.77
D268	311.73	318.98	-7.25	PVGPU01	402.53	406.73	-4.20
D282	309.15	310.30	-1.15	PVGSS05	434.38	449.05	-14.67
D298	309.06	312.93	-3.87	PVMSW01	410.12	428.62	-18.50
D301	309.26	312.54	-3.28	PVMSW04	370.25	384.23	-13.98
D400	326.21	313.02	13.19	PVRAB05	423.84	435.98	-12.14
D402	363.29	348.89	14.40	PVVCA05	307.85	301.81	6.04
D404	336.24	341.98	-5.74	PVVCA06	300.92	299.88	1.04
D412	414.46	415.95	-1.49	PVWBG12	409.77	426.83	-17.06
D413	320.70	326.11	-5.41	PVWCC15	334.89	345.00	-10.11
D414	315.80	326.53	-10.73	PVWCV05	346.02	358.59	-12.57
D435	289.56	289.25	0.32	PVWLY07	412.73	409.19	3.54

Table 7b. Calculated and simulated ground-water discharge to surface water, Pohatcong Valley, Warren County, N.J.

[Streamflow was measured August 8, 2000; ft^3/s, cubic feet per second; Cr., Creek; Trib., tributaries; Number in parenthesis is station number; --, not calculated]

Name of stream reach	Calculated net gain (ft^3/s)	Simulated net gain (ft^3/s)	Residual (measured - simulated) (ft^3/s)	Coefficient of variation
Pohatcong Cr. at Mine Hill Rd. (01455140)	1.41	1.54	0.133	0.47
Pohatcong Cr. at Rt. 57 (01455145)	1.74	1.09	-0.65	0.472
Pohatcong Cr. at Brass Castle Cr. (01455155)	1.58	1.64	0.058	0.642
Pohatcong Cr. at Broadway (01455180)	1.90	2.79	0.891	0.569
Pohatcong Cr. at New Village (01455200)	-0.05	0.32	0.368	25.5
Pohatcong Cr. at Carpentersville (01455300)	16.72	19.21	2.477	0.311
Shabbecong Cr.	0.19	0.23	0.036	0.178
Brass Castle Cr.	-0.68	-0.33	0.351	0.312
North Side Tributaries (Trib. 7, 8, 10)	-0.04	0.07	0.111	1.39
South Side Tributaries (Trib. 9 and adjacent)	0.18	0.00	-0.178	0.149
Quarry	4.50	2.34	-2.158	--

charge to the quarry were considered qualitatively with the goal of avoiding any gross errors.

Simulated Ground-Water Age

The age of ground water was simulated by calculating the time of travel along the simulated flow paths from each of the 10 bedrock wells (for which samples were age dated) back to the point of recharge. The ages determined from SF6, CFC, or tritium-helium measurements range from 0 to 27 years (table 8). Porosity was the only parameter that was adjusted to calibrate simulated travel times. This porosity represents the bulk porosity of the formation and does not account for the wide variations in porosity caused by the presence of solution cavities or unfractured rock. The wide range in measured ages and inherent uncertainties of age-dating techniques, especially in industrialized areas, allow only an order of magnitude estimate of porosity. Porosities of 0.2 (20 percent) for the surficial (unconsolidated) aquifer and 0.01 (1 percent) for each of the layers representing the carbonate aquifer yielded median travel times to the 10 bedrock wells of 2 to 31 years (table 8), which was considered to be a good match.

Model limitations

Because the flow of ground water in a fractured carbonate-rock environment is too complex to be precisely represented with a numerical model, many simplifying assumptions must be made, especially where model-input data are sparse or nonexistent. For example, wide variations in horizontal hydraulic conductivity were measured and these heterogeneities are important to the flow system, but were beyond the scope of this project to simulate explicitly; therefore, averaged values were used. In addition, the ground-water-flow model was designed to simulate flow on a regional scale. Caution was used when evaluating results (for example, simulated ground-water-flow paths) at a specific location because actual conditions could vary substantially from simulated conditions as a result of inherent model uncertainties and the effects of heterogeneities that are not accounted for in the model. The combination of model parameters that were selected to achieve the best fit to measured water levels and base flows is not unique. Different combinations of the parameters could produce similar overall fit to the measured data but different flow paths or other results. The application and accuracy of the conceptual and numerical models are limited by uncertainties, for example, uncertainties in the hydraulic properties of the carbonate-rock aquifer near the valley wall. The high water levels along the valley walls were not well reproduced in the model. Some of the wells identified as being open to the carbonate-rock aquifer may be in fact be open to the crystalline-rock aquifer, or the variation in hydraulic conductivity may be larger than simulated. The model is calibrated to stream base flow and water-level data that were collected in different years. Furthermore, neither data set exactly represents mean annual conditions.

Table 8. Results of age dating of ground-water samples using sulfur hexafluoride (SF_6), chlorofluorocarbon (CFC), and tritium-helium isotope (H^3-He^3) techniques, Pohatcong Valley, Warren County, N.J.

[--, not measured; Contaminated, indicates tracer gas concentrations were higher than northern hemisphere average because of a local source]

Well Identifier	SF_6 age, in years	CFC age, in years	H^3-He^3 age, in years	Median simulated age of particles reaching model cell in which sampled well is located
D206	6	Contaminated	--	13
D211	0 - 2	--	--	31
PVANC06	27	Contaminated	6	4
PVBMC01	7	--	--	9
PVFCC12	12	Contaminated	21	12
PVFCC13	4	Contaminated	7	--
PVGPU01	11	15	--	6
PVSTJ01A	4	--	--	8
PVWCC14	Contaminated	0-18	--	--
PVWCC15	3	0 - 2	--	8
PVWCV05	5	Contaminated	9	13
PVWLY08	12	--	--	5

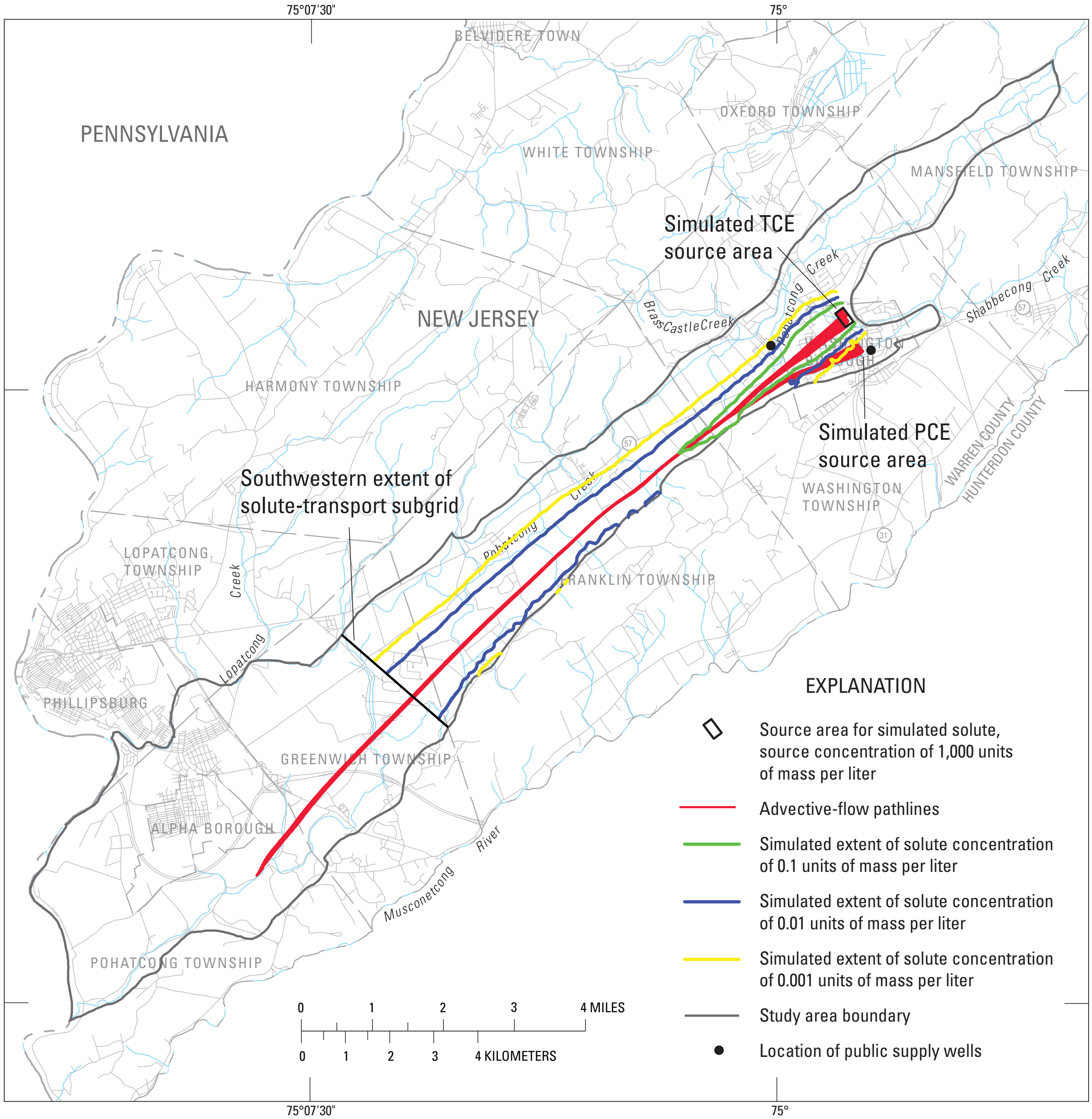
Simulated Ground-Water-Flow Paths

Ground-water-flow paths beginning or ending at any location can be simulated using the USGS MODPATH particle-tracking package of MODFLOW (Pollock, 1994). In order to visualize flow paths in the Pohatcong Valley, particles were started in the carbonate-rock aquifer in the TCE and PCE source areas identified by CH2M Hill (2005a) (fig. 21). Particles originating in the TCE source area proceed downvalley, exiting the model domain at Pohatcong Creek. Some flow paths end just above or below the confluence of Pohatcong and Shabbecong Creeks, but some go deeper into the aquifer and discharge to Pohatcong Creek nearly 10 miles downvalley. Particles originating in the PCE source area flow parallel to Shabbecong Creek and discharge to Pohatcong Creek just below the confluence with Shabbecong Creek. When the modeled PCE source area was extended about 300 feet to the southeast, a few particles were captured by production well PVMSW01 (fig. 1).

The advective flow paths shown in figure 21 are relatively insensitive to changes in most model parameters. In the model, the main influence on the length of the flow path is the rate of ground-water discharge to Pohatcong Creek in the vicinity of the confluence with Shabbecong Creek. When model parameters were adjusted such that simulated base flows were substantially higher than measured base flows in the reach of Pohatcong Creek from surface-water station 01455140 to station 01455145 (near Rt. 57, just upstream from the confluence with Shabbecong Creek) and the reach between the confluences with Shabbecong and Brass Castle Creeks (fig. 8), all pathlines originating at the TCE and PCE source areas ended in these reaches. The pathlines shown in figure 21 represent estimated average annual conditions; therefore, periods of high ground-water levels could cause more ground-water discharge to Pohatcong Creek upstream from the confluence with Brass Castle Creek, whereas periods of low water levels could cause Pohatcong Creek to recharge the aquifer, resulting in the flow paths extending much farther downvalley.

The long, narrow advective-flow pathlines shown in figure 21 do not account for the dispersive nature of solute transport. Although the largest mass of solutes generally will follow the advective flow path, dispersion will increase the width of a solute plume. Solute transport was simulated by adding the USGS ground-water-transport process MOC3D (Konikow and others, 1996) to the numerical model. Additional parameters included in the model were porosity and dispersivity. The surficial-aquifer porosity was set to 0.2, and the carbonate-rock-aquifer porosity was set to 0.01. The longitudinal, transverse, and vertical dispersivities were set to 1, 1, 0.1 m, respectively, for the surficial aquifer and 75, 7.5, and 0.75 m, respectively, for the carbonate-rock aquifer. A solute was introduced to the shallowest carbonate-rock-aquifer model layer, injecting into each of six model cells in the TCE source area 100 liters of water per day containing 1,000 units of mass per liter of solute. The simulation was run for 10 years.

The solute-transport simulation was not calibrated to field data, and is for illustrative purposes only, but does show a plume extending down the valley with a greater width than the advective flow pathlines (fig 21). Although the units of solute concentration are arbitrary, 1,000 units per liter is within a factor of two of the solubility constant of TCE



Base from U.S. Geological Survey digital line graph files, 1:24,000

Figure 21. Simulated advective-flow pathlines of particles originating in the trichloroethene (TCE) and tetrachloroethene (PCE) source areas and simulated extent of selected concentrations of a solute originating from a constant source in the vicinity of the TCE source area, Pohatcong Valley, Warren County, N.J.

in water (measured in milligrams per liter). Similarly, the maximum solute concentration observed in the model, 2 units of mass per liter, also is within a factor of two of the highest measured concentration of TCE (measured in milligrams per liter) in the carbonate-rock aquifer (CH2M Hill, 2005a), and the downvalley and lateral extents of the simulated 0.1 unit of mass-per-liter contour and lateral extents of the 0.01 and 0.001 contours (fig. 21) are similar to the extents of those order of magnitude concentrations (in milligrams per liter) measured in the field (CH2M Hill, 2005a). However, the simulation does not include the heterogeneities (other than those described previously) of the aquifers. Unlike the simulated concentration distributions, actual contaminant distributions are greatly affected by small-scale and large-scale heterogeneities, such as fracture distributions, solution channels, and faults.

Simulation of Ground-Water Remediation Alternatives

Four ground-water remediation alternatives have been proposed for the USEPA Pohatcong Valley Ground Water Contamination Site (CH2M Hill, 2005b). These include GW1, no further action; GW2, source-area withdrawal, treatment, and reinjection with pumping rates of 420 gal/min in the TCE source area and 100 gal/min in the PCE source area; GW3, expanded source-area withdrawals, treatment, and reinjection with pumping rates of 1,400 and 420 gal/min in the TCE and PCE source areas, respectively; and GW4, entire plume collection and treatment with pumping rates of 1,400 gal/min in the TCE source area and 7,500 gal/min along the entire length of the TCE plume. Alternative GW1 represents current conditions (fig. 21), the simulation of which is described in preceding sections of this report. Simulated flow paths for alternatives GW2, GW3, and GW4 are shown in figures 22-24. For the alternatives GW2, GW3, and GW4, the intended source area can be designated and particles originating in the source area tracked in the forward direction to see if they are captured by the withdrawal wells. Particles originating in the simulated source areas are shown as captured or not captured under the different pumping alternatives in figures 22 to 24. The source area represented as the starting locations of particles is limited to the outlines of model cells and, therefore, is not a smooth curve but rather an angular outline of the included model cells. Given the incomplete knowledge of the exact location of the source areas, the blocky outline of the simulated source area is not considered unreasonable.

Alternative GW2

Ground-water alternative GW2 was designed to capture those parts of the plumes with the highest concentrations of TCE and PCE (CH2M Hill, 2005b). The intended capture area for the TCE plume is the area within the 500-μg/L isopleth (CH2M Hill, 2005a, figs. 5 and 6). The designated area is somewhat irregular because of the heterogeneity of the system. The intended capture zone for the PCE plume was not explicitly identified by CH2M Hill (2005b) because of the paucity of data. CH2M Hill (2005b) proposes installing three withdrawal wells in the TCE plume, each open to the aquifer from 110 to 310 feet below land surface (ft bls) and pumping at 140 gal/min, and one withdrawal well in the PCE plume, open from 80 to 280 ft bls and pumping at 100 gal/min. After treatment, the withdrawn water would be reinjected into the aquifer through four injection wells upgradient from the TCE source area (each open from 110 to 310 ft bls, injecting 105 gal/min) and one injection well upgradient from the PCE source area (open from 80 to 280 ft bls, injecting 100 gal/min).

The pathlines of particles originating in the TCE source area (within the discretized 500-μg/L isopleth) and the PCE source area are shown in figures 22a and 22b. The location of the withdrawal and injection wells in the PCE source area and the injection wells near the TCE source area are the same in both figures. The locations of the withdrawal wells in the TCE source area in figure 22a are those proposed by CH2M Hill (2005b), but the wells do not capture all particles originating within the discretized 500 μg/L isopleth. A slight realignment of the wells to a line more nearly perpendicular to the direction of anisotropy simulated in the finite-difference model captures all of the particles originating within the discretized TCE 500 μg/L isopleth (fig. 22b). The discretized PCE source area is fairly large for the withdrawal rate proposed for the lone withdrawal well; therefore, only a small subset of particles would be captured by the GW2 withdrawal well.

Alternative GW3

Ground-water alternative GW3 is designed to capture a larger part of the TCE and PCE plumes than GW2 (CH2M Hill, 2005b). The intended capture zone for the TCE plume is the area within the 100-μg/L isopleth in the vicinity of the TCE source area. The intended capture area for the PCE plume was not explicitly identified by CH2M Hill (2005b) because of the paucity of data. CH2M Hill (2005b) proposes five withdrawal wells in the TCE source area, each open to the aquifer from 110 to 310 ft bls and pumping at 280 gal/min (total withdrawal rate of 1,400 gal/min), and two withdrawal wells in the PCE source area, open from 80 to 280 ft bls and pumping at 210 gal/min (total withdrawal rate of 420 gal/min). After treatment, the withdrawn water would be reinjected into the aquifer through eight injection wells upgradient from the TCE source area (open from 110 to 310 ft bls, injecting 175 gal/min) and four injection wells near the PCE source area (open from 80 to 280 ft bls, injecting at 105 gal/min).

The pathlines of particles originating in the TCE plume (within the discretized 100-μg/L isopleth) and the discretized PCE source area are shown in figures 23a and 23b. The location of the withdrawal and injection wells in the PCE source area and the injection wells near the TCE source area are the

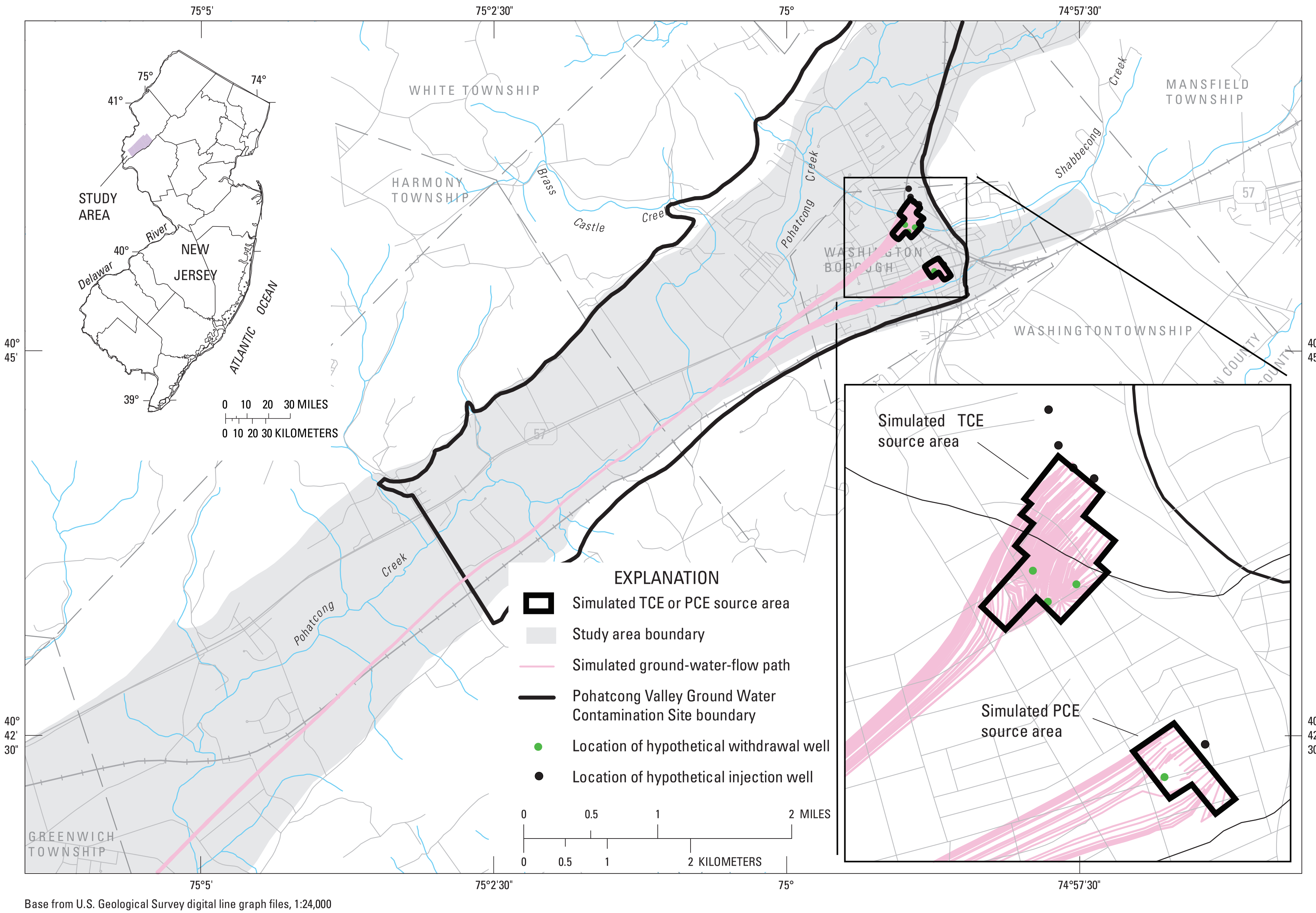
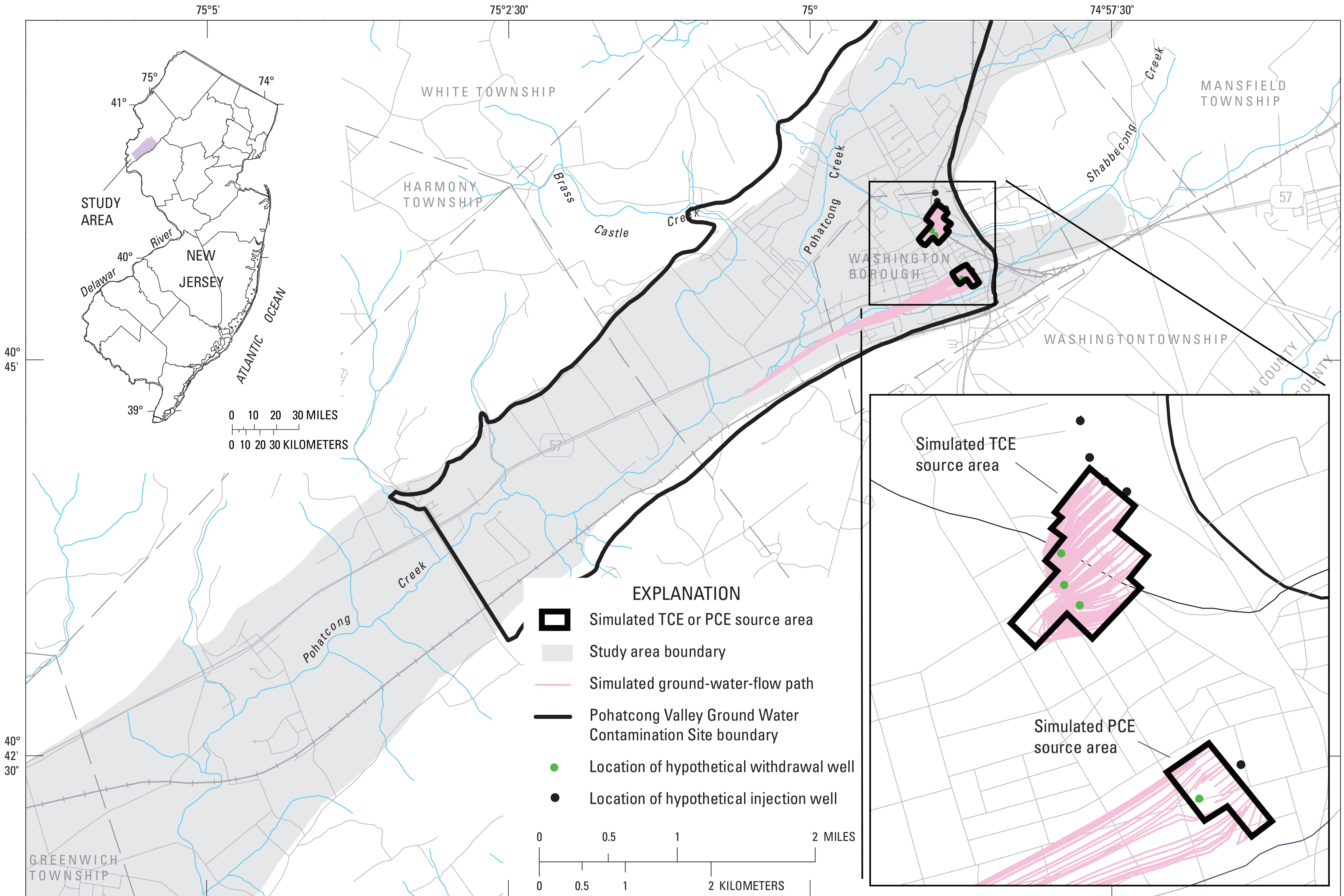


Figure 22a. Simulated ground-water-flow paths originating from the tetrachloroethene and trichloroethene (concentration greater than 500 micrograms per liter) source areas, assuming implementation of ground-water remediation alternative GW2, Pohatcong Valley, Warren County, N.J. (PCE, tetrachloroethene; TCE, trichloroethene; see text for description of GW2)



Base from U.S. Geological Survey digital line graph files, 1:24,000

Figure 22b. Simulated ground-water-flow paths originating from the tetrachloroethene and trichloroethene (concentration greater than 500 micrograms per liter) source areas, assuming modified implementation of ground-water remediation alternative GW2, Pohatcong Valley, Warren County, N.J. (PCE, tetrachloroethene; TCE, trichloroethene; see text for description of GW2) same in figures 23a and 23b. The location of the withdrawal wells in the TCE plume in figure 23a are those proposed by CH2M Hill (2005b). The withdrawal wells capture all particles originating within the discretized 100-μg/L isopleth upgradient from the withdrawal wells and for up to about 1,000 ft downgradient. However, the large pumping rate would allow a larger capture zone. Another simulation was made with the withdrawal wells moved farther from the TCE source area. In this simulation, particles originating in a larger part of the source area are captured (fig. 23b). The same source area for the PCE plume was used in figures 23a and 23b as was used in figures 22a and 22b. One well withdrawing at 100 gal/min was not sufficient to capture all of the particles emanating from the discretized PCE source area, but two wells withdrawing at a total rate of 420 gal/min do capture almost all of the particles. A few particles on the ground-water divide between the capture zones of PVMSW01 and the GW3 withdrawal wells are not captured, but this is not considered important.

Alternative GW4

Ground-water alternative GW4 is designed to capture the entire TCE and PCE contaminant plume (concentration greater than 1 μg/L) in the Pohatcong Valley (CH2M Hill, 2005b). The withdrawal and injection wells proposed in the GW3 alternative (pumping at the rate of 1,820 gal/min) are augmented by six sets of five withdrawal wells, with each set of wells spaced approximately 4,000 feet apart along the longitudinal extent of the valley. The five withdrawal wells in each set, each pumping 300 gal/min (1,500 gal/min per set, 9,000 gal/min total), are spread laterally across the valley, spanning the width of the plume (concentrations greater than 1 μg/L). CH2M Hill (2005b) does not propose a location for sets of injection wells and extends the six sets of withdrawal wells only to the end of the Pohatcong Valley Ground Water Contamination Site boundary (fig. 1). For this study, the spacing between each set of withdrawal wells was increased so that the set farthest downvalley is just below the currently known (2005) extent of TCE contamination, about 2 miles southwest of New Village. Each set of five injection wells was arbitrarily placed about 1,000 ft downvalley of the associated set of withdrawal wells, except for the southwestern-most set, which was placed upvalley from a residential area so as to flush the contamination out of that residential area more quickly and allow the last set of withdrawal wells to capture the southwestern extent of known (2005) contamination. The total withdrawal rate for the GW4 system shown in figure 24 is 10,820 gal/min (15.6 Mgal/d).

The pathlines of particles originating in a subset of the known (2005) extent of TCE and PCE contamination exceeding 1 μg/L are shown in figure 24. A map showing pathlines of particles originating throughout the entire extent of the contaminated area was not legible. Therefore, to illustrate the behavior of the system, one set of particles originates in the area within the discretized 1-μg/L isopleth from the TCE and PCE source areas to about 1,000 ft upgradient from the first set of five withdrawal wells located laterally across the valley. The second set of particles originates in a 2,000-ft wide area downvalley from the second lateral set of injection wells and upvalley from the third set of withdrawal wells. Most of the particles originating in the two subsections of the contaminant plume are captured by the remedial system (fig. 24). Some of the particles travel deep enough in the modeled aquifer that they are not captured by the withdrawal wells located in the top 300 ft of the aquifer.

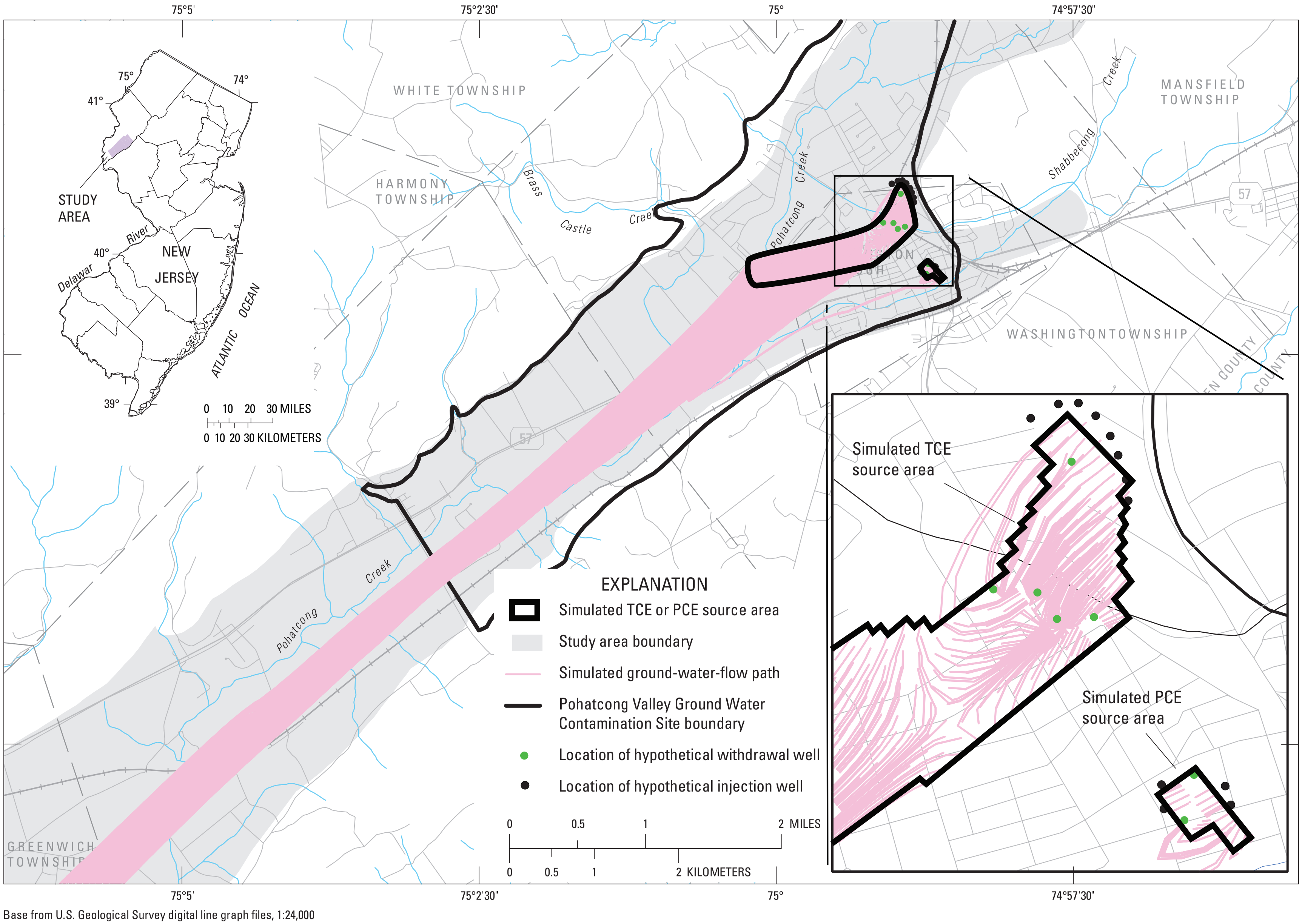


Figure 23a. Simulated ground-water-flow paths originating from the tetrachloroethene and trichloroethene (concentration greater than 100 micrograms per liter) source areas, assuming implementation of ground-water remediation alternative GW3, Pohatcong Valley, Warren County, N.J. (PCE, tetrachloroethene; TCE, trichloroethene; see text for description of GW3)

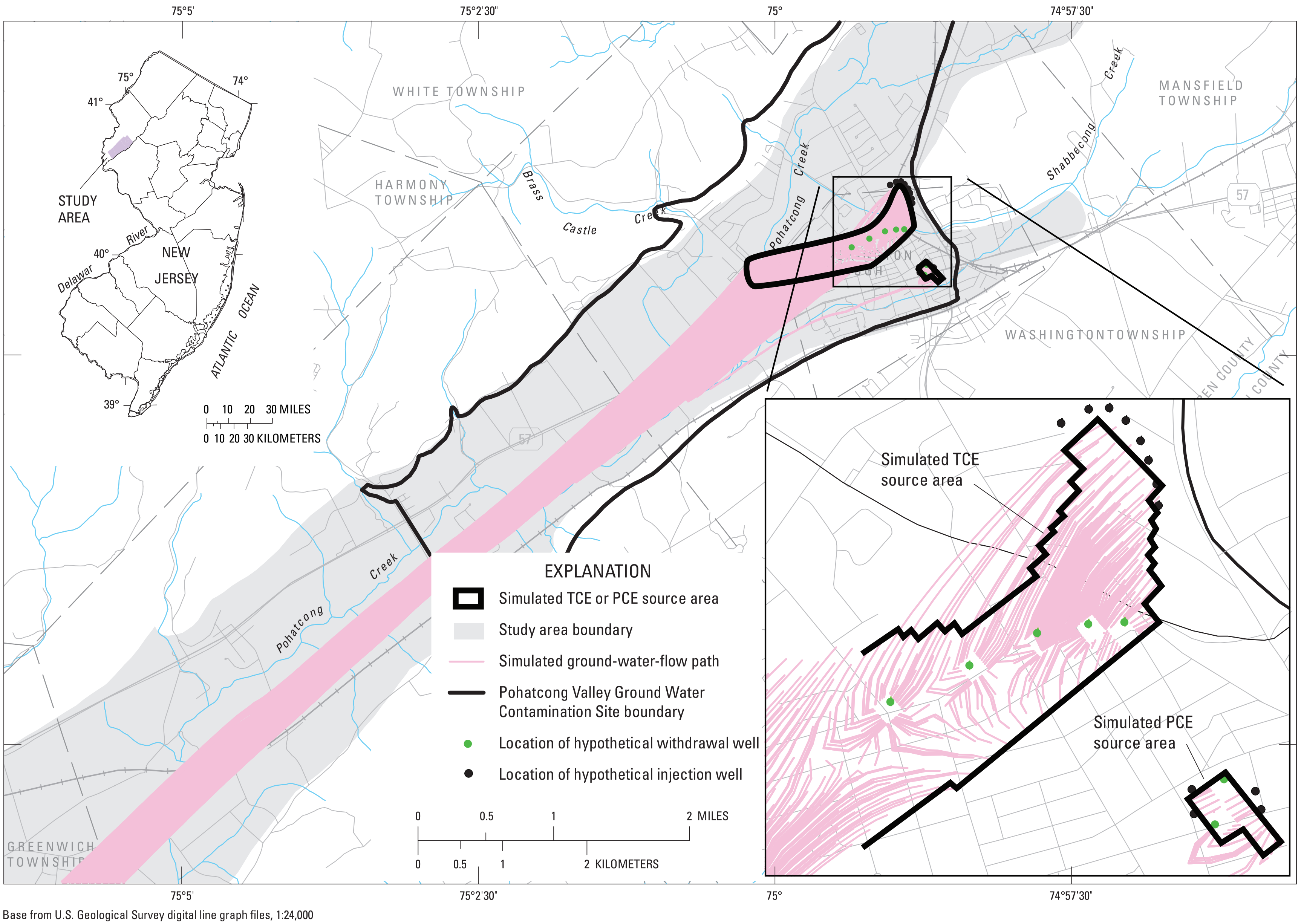
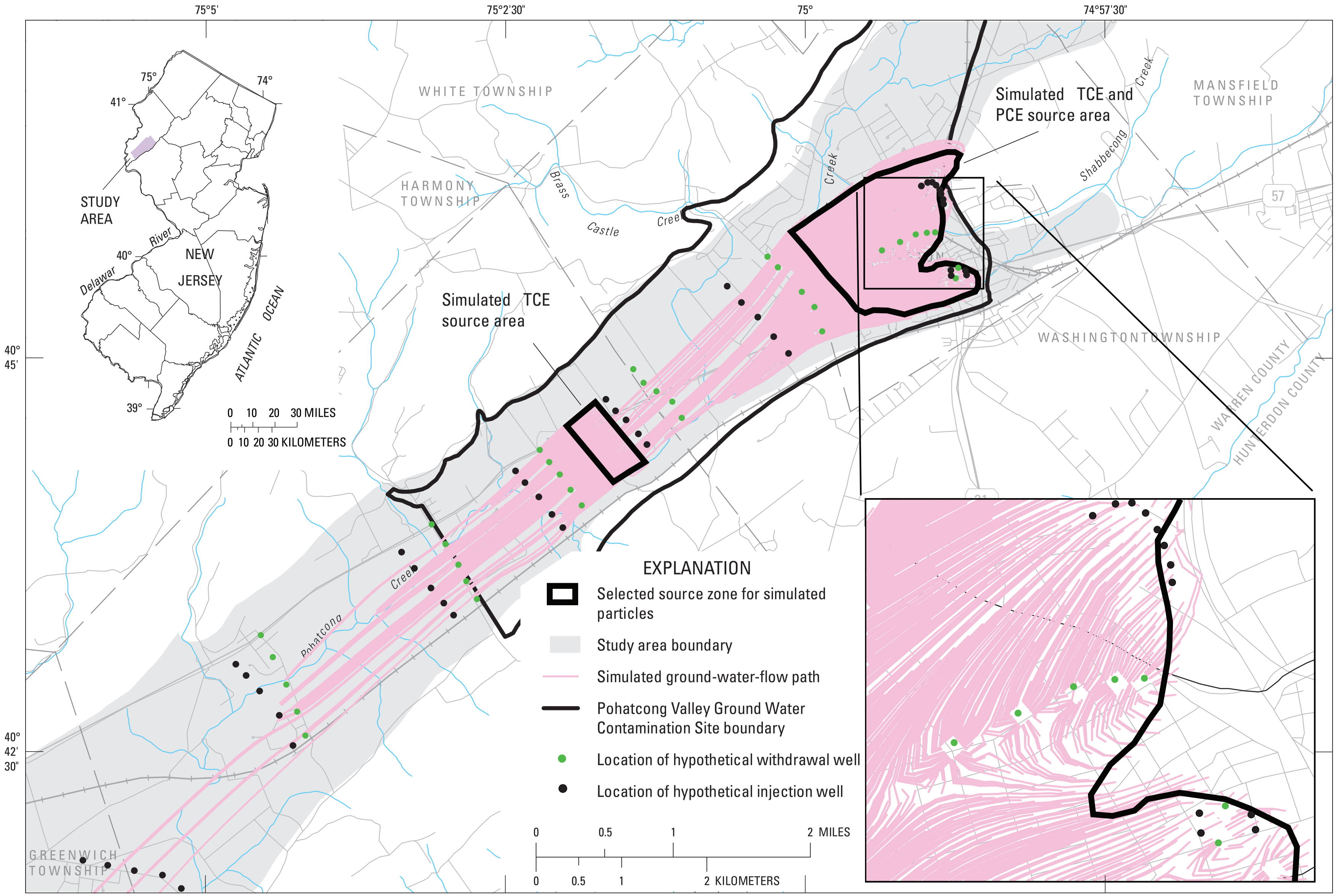


Figure 23b. Simulated ground-water-flow paths originating from the tetrachloroethene and trichloroethene (concentration greater than 100 micrograms per liter) source areas, assuming modified implementation of ground-water remediation alternative GW3, Pohatcong Valley, Warren County, N.J. (PCE, tetrachloroethene; TCE, trichloroethene; see text for description of GW3)



Base from U.S. Geological Survey digital line graph files, 1:24,000

Figure 24. Simulated ground-water-flow paths originating from the tetrachloroethene source area and selected locations in the trichloroethene (concentration greater than 1 microgram per liter) source areas, assuming implementation of ground-water remediation alternative GW4, Pohatcong Valley, Warren County, N.J. (PCE, tetrachloroethene; TCE, trichloroethene; see text for description of GW4)

Summary

The chlorinated solvents trichloroethene (TCE) and tetrachloroethene (PCE) were detected in 1978 in production wells in Washington Borough and Washington Township, Pohatcong Valley, Warren County, New Jersey. Subsequent investigation revealed that many domestic wells in Washington and Franklin Townships also were contaminated, and in 1989 the Pohatcong Valley Ground Water Contamination Site was added to the U.S. Environmental Protection Agency (USEPA) National Priority List. A remedial investigation by the USEPA and CH2M Hill was begun in 1999. The U.S. Geological Survey (USGS) provided technical assistance to the USEPA with various hydrogeologic aspects of the remedial investigation and constructed a numerical model to simulate various aspects of the ground-water-flow system of the Pohatcong Valley. The analyses and simulations described in this report were conducted by the USGS in cooperation with the USEPA.

The Pohatcong Valley is underlain by surficial deposits and carbonate bedrock and is bordered by crystalline rocks. The surficial deposits form a minor aquifer, the thickness of which is highly variable, partly because of localized extensive weathering of the underlying carbonate rocks. Although five different carbonate formations have been identified within the Pohatcong Valley, the formations can be grouped together and considered a single carbonate-rock aquifer. The surficial aquifer receives recharge from direct infiltration of precipitation and runoff from the crystalline-rock areas. Ground-water flow in the carbonate rocks is generally downvalley towards the Delaware River, although there is discharge through the surficial aquifer to Pohatcong Creek under average conditions.

The hydraulic characteristics of the carbonate-rock aquifer are highly heterogeneous. Fracture data indicate that water-producing fractures occur in most orientations, which supports the use of a finite-difference model to simulate the aquifer as an equivalent porous media. There are slightly more fractures striking approximately northeast to southwest, supporting the conclusion from aquifer test data that the aquifer is horizontally anisotropic. Results of 47 hydraulic tests using packers to isolate short sections of open boreholes show hydraulic conductivities that span nearly five orders of magnitude, from 0.5 ft/d to 1,800 ft/d. The maximum transmissivity calculated was 37,000 ft^2/d. Specific capacity data from four carbonate formations and the adjacent crystalline rock indicate the carbonate formations have similar permeabilities, supporting the approach of modeling the carbonate formations as a single aquifer (without significant intervening confining units). Specific-capacity data from 42 wells completed in the carbonate rock and 32 wells completed in Precambrian crystalline rock indicate the crystalline rock is at least an order of magnitude less permeable than the carbonate rock, supporting the approach of treating the crystalline rock as a no-flow boundary in the model. The median horizontal hydraulic conductivity calculated from specific capacity data from 20 surficial-aquifer wells is 2.7 ft/d. A large-scale aquifer test conducted using a production well open to the Leithsville Formation yielded a hydraulic conductivity value of 34 ft/d. A large-scale aquifer test using a production well open to the Allentown Dolomite resulted in the estimated horizontal anisotropy of 0.46 and maximum and minimum calculated horizontal hydraulic conductivities of 190 and 85 ft/d, respectively.

Stream base-flow data were used to estimate the net gain (or loss) for five reaches on Pohatcong Creek, and for Brass Castle Creek, Shabbecong Creek, and three smaller tributaries. Hydrographs created from historical data from former continuous-record stations in the valley and from nearby current continuous-record stations were used to separate streamflow into base flow and runoff components. Statistical correlations between three partial-record stations in Pohatcong Valley and four nearby continuous-record stations made possible the calculation of estimated mean annual base flows for the three partial-record stations. Estimated mean annual base flows for Brass Castle Creek, Pohatcong Creek at New Village, and Pohatcong Creek at Carpentersville are 2.4, 25, and 45 ft^3/s (10, 10, and 11 in/yr), respectively.

Ground-water ages were estimated by evaluating concentrations of trace dissolved atmospheric gases sulfur hexafluoride (SF_6), chlorofluorocarbons (CFCs), and tritium-helium in samples from 12 wells. Samples from most of the wells had CFC concentrations exceeding atmospheric levels and 1 well had SF_6 concentrations exceeding atmospheric levels, indicating contamination from a nearby source. Ground-water ages could not be estimated from CFC or SF_6 concentrations if contamination was present. The ground-water ages determined from SF_6 and CFC concentrations for the remaining samples ranged from 27 years to modern (0-2 years), with a median age of 6 years. Ages determined for three of the four ground-water samples using the tritium-helium technique ranged from 7 to 21 years and were nearly twice the ages determined using SF_6. The age determined for the fourth ground-water sample using the tritium-helium technique (6 years) was less than a quarter of the age determined using the SF_6 technique (27 years). Despite the uncertainties, the results of age dating clearly indicate that ground water in the valley moves through the system rapidly and is unlikely to be more than about 25 years old.

Estimated ground-water and land-surface water budgets were developed. The land-surface water budget was used to estimate the rate of direct recharge to the surficial aquifer. The estimated rate of direct recharge to the surficial aquifer underlain only by the carbonate-rock aquifer is about 23 in/yr. The ground-water budget accounts for water entering the ground-water system through all sources of recharge (inputs) and leaving the system as ground-water discharge (outputs). The estimated net recharge to the ground-water system within the area underlain by carbonate rock (11.4 mi^2) is about 29 in/yr (the net ground-water recharge for the entire basin—crystalline and carbonate-rock areas—is 10 in/yr).

A finite-difference, numerical model was developed to simulate ground-water flow in the Pohatcong Valley. The model encompasses the entire carbonate-rock part of the val-

ley, from near the headwaters of Pohatcong Creek to the Delaware River and has four layers; layer 1 represents the surficial aquifer, and layers 2-4 represent equal-thickness parts of the carbonate-rock aquifer. Layer 1 was modeled as unconfined, and layers 2-4 were modeled as confined, although for some model runs layer 1 (the surficial aquifer) was modeled as confined. When the surficial aquifer was modeled as unconfined, many cells 400 or more feet from Pohatcong Creek, especially in the southwestern half of the valley, dried up and were inactive. Changing layer 1 from confined to unconfined did not produce important differences in model results. The carbonate-rock aquifer was modeled as horizontally anisotropic, with the direction of maximum transmissivity aligned with the longitudinal axis of the valley.

The lateral boundaries of the model were all no-flow boundaries. Although the Delaware River to the south and Lopatcong Creek on the northwestern edge of the southwestern end of the modeled area technically were represented with overlying head-dependent boundaries, they functioned as lateral boundaries. All surface-water bodies, including Pohatcong Creek, its tributaries, and an abandoned, water-filled quarry, were represented with head-dependent boundaries. Pohatcong Creek and all tributaries except Brass Castle Creek were represented as drains, meaning they act as a boundary only if head in the underlying aquifer is higher than the stage of the overlying boundary. Recharge was applied uniformly to the topmost active layer, except near the lateral boundaries where additional recharge was added to represent infiltration of runoff from adjacent crystalline-rock areas.

The model was calibrated to water levels in the carbonate-rock aquifer measured in June 2001 and stream base flow measured August 8, 2000. Aquifer porosities were adjusted (using the previously calibrated model) until the approximate average age of particles traveling to the sampled wells was about 5 years. The final calibrated maximum horizontal hydraulic conductivities for layers 1 through 4 were 1.8, 300, 100, and 33 ft/d; vertical hydraulic conductivities were 6.6, 100, 33, and 10 ft/d. Horizontal and vertical hydraulic conductivities in model layers 2, 3, and 4 are 0.08 times lower than the maximum in areas representing the Leithsville Formation, 0.7 times lower than the maximum in areas representing the Beekmantown Group and Jacksonburg Limestone, and 0.018 times lower around the perimeter. Horizontal anisotropy of the carbonate rock was 0.2, and porosity was 0.01 (1 percent). Streambed hydraulic conductivities ranged from 0.13 to 26 ft/d for Pohatcong Creek and its tributaries, 0.001 ft/d for the Delaware River (modeled as directly connected to the bedrock aquifer), and 660 ft/d for the quarry. Recharge was set to a total of 45 ft^3/s (11 in/yr if applied over the entire basin, 32 in/yr over the carbonate-rock part of the basin).

Four ground-water remediation alternatives (GW1, GW2, GW3, and GW4) described by CH2M Hill were simulated. GW1 is the no-action alternative; therefore, output from the simulation represents current (2000-2005) conditions and shows pathlines that originate in the PCE source area (in central Washington Borough) and flow southwest until they end at Pohatcong Creek near the confluence with Shabbecong Creek. Most pathlines originating in the TCE source area (at the northern edge of Washington Borough) end at Pohatcong Creek near the confluence with Shabbecong Creek, although some pathlines go deeper in the aquifer system and ultimately end at Pohatcong Creek about 10 miles downvalley in Pohatcong Township.

Alternatives GW2, GW3, and GW4 include ground-water withdrawals, treatment, and reinjection. The design for GW2 includes three wells in the TCE source area that withdraw water at a total rate of 420 gal/min and one well in the PCE source area withdrawing water at a rate of 100 gal/min. Depending on placement of the three wells in the TCE source area, flow-path analysis shows the system would capture all ground water within the discretized 500-μg/L TCE isopleth. The single well in the PCE source area would capture water from only a small area.

The design for GW3 includes five wells in the TCE source area withdrawing water at a total rate of 1,400 gal/min and two wells in the PCE source area withdrawing water at a total rate of 420 gal/min. Flow-path analysis shows the system would capture all ground water within the discretized 100-μg/L TCE isopleth in the source area and some of the ground water within the 100-μg/L TCE isopleth downgradient from the source area. The system would capture all of the ground water in the estimated PCE source area.

The design for GW4 includes the GW3 wells withdrawing water at a total rate of 1,820 gal/min plus six sets of withdrawal wells placed at 4,000-ft intervals down the valley, with five wells per set (30 wells total) withdrawing a total of 9,000 gal/min (total withdrawal rate of 10,820 gal/min). Most particles started in the TCE source area and in an arbitrary area representing contamination farther downvalley were captured by the GW4 system, although a few pathlines traveled beneath the withdrawal wells and down the valley, ultimately ending at Pohatcong Creek.

Acknowledgments

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Table 1. Location and well construction information for selected wells, Pohatcong Valley, Warren County, New Jersey.

[USGS, U.S. Geological Survey; NJDEP, New Jersey Department of Environmental Protection; USEPA, U.S. Environmental Protection Agency; 112SFDF, Stratified drift; 360KTTN, Kittatiny Group, 371ALNN, Allentown Formation; 374LSVL, Leithsville Formation; 400PCMB, Precambrian gneiss; lat, latitude; long, longitude; NAD27, North American Datum of 1927; NAD83, North American Datum of 1983; NGVD29, National Geodetic Vertical Datum of 1929; NAVD88, North American Vertical Datum of 1988; M, map; G, Global Positioning System; L, Leveled; D, Differential Global Positioning System; --, no data or not applicable]

Well identification code	USGS well number	NJDEP permit number	USEPA well number	Local well number or name	Aquifer code	Latitude (NAD83)	Longitude (NAD83)	Lat/Long converted from NAD27 to NAD83	Method used to determine lat/long	Land surface altitude (feet above NAVD88)	Altitude converted from NGVD29 to NAVD88	Method used to determine altitude	Estimated accuracy of altitude (feet)	Depth of well (feet)	Top of open interval below land surface (feet)	Bottom of open interval below land surface (feet)
A3	410422	24-24483	PV095DW	DOM A3	371ALNN	404650.29	0745848.46	--	G	457.55	--	L	0.1	200	67	200
A470	410486	--	--	A470	360KTTN	404410.93	0750314.76	--	G	398.3	--	D	.3	86.6	--	--
A472	410482	--	--	A472	360KTTN	404340.77	0750352.40	--	G	386.4	--	D	.3	102	--	--
A474	410557	--	PVDOM02	PVDOM02 (A474)	371ALNN	--	--	--	G	406.8	--	L	.3	112	55	112
C488	410488	--	--	PS C488	360KTTN	404412.23	0750126.90	--	G	375.4	--	D	.3	25	--	--
D55	410421	24-24417	--	DOM D55	400PCMB	404648.68	0745813.26	--	G	493	--	M	10	124	102	124
D57	410419	24-25036	PV088DW	DOM D57	371ALNN	404632.52	0745953.34	--	G	474	--	M	10	248	131	248
D58	410418	24-25892	--	DOM D58	400PCMB	404621.97	0750005.04	--	G	509	--	M	10	445	60	445
D60	410423	24-25779	--	DOM D60	374LSVL	404656.32	0745746.59	--	G	508.6	--	L	.1	123	100	123
D70	410420	24-31166	--	DOM D70	374LSVL	404645.87	0745836.27	--	G	455.2	--	L	.1	165	118.5	165
D71	410417	24-30778	--	DOM D71	400PCMB	404611.64	0750010.69	--	G	527	--	M	10	265	161	265
D84	410558	24-19372	PV090DW	PV090 (D84)	400PCMB	404600.32	0750031.61	--	G	545	--	M	10	150	86	150
D206	410397	24-23773	--	DOM D206	371ALNN	404307.4	0750457.19	--	G	426.0	--	L	.1	172	170	172
D211	410393	24-27055	PV079DW	DOM D211	360KTTN	404246.86	0750416.7	--	G	367.0	--	L	.1	123	121	123
D218	410402	24-28174	--	DOM D218	371ALNN	404321.3	0750419.52	--	G	406.7	--	L	.1	168	148.5	168
D225	410394	24-29134	--	DOM D225	360KTTN	404259.4	0750426.52	--	G	326.9	--	L	.1	185	159.5	185
D234	410395	24-31780	--	DOM D234	360KTTN	404305.58	0750436.23	--	G	386.6	--	L	.1	305	103.5	305
D238	410392	24-33224	--	DOM D238	360KTTN	404241.26	0750414.17	--	G	381.7	--	L	.1	205	183.5	205
D242	410396	24-33503	--	DOM D242	371ALNN	404305.84	0750506.31	--	G	391.9	--	L	.1	148	140	148
D244	410391	24-33589		DOM D244	360KTTN	404237.51	0750433.92	--	G	381.4	--	L	.1	205	153.5	205
D246	410398	24-34341	--	DOM D246	360KTTN	404309.25	0750442.78	--	G	406.7	--	L	.1	165	124.5	165
D268	410403	24-15310	--	DOM D268	371ALNN	404333.32	0750410.34	--	G	414.8	--	L	0.1	185	135	185
D282	410399	24-09132	PV076DW	DOM D282	360KTTN	404313.3	0750428.22	--	G	371.2	--	L	.1	103	74	103
D298	410401	24-16331	--	DOM D298	371ALNN	404319.01	0750425.58	--	G	400.4	--	L	.1	166	115	166
D301	410400	24-16735	--	DOM D301	360KTTN	404316.97	0750422.36	--	G	404.2	--	L	.1	225	168	225
D400	410479	--	--	DOM D400	360KTTN	404307.13	0750505.60	--	G	391.6	--	D	.3	145.6	--	--
D403	410484	--	--	DOM D403	112SFDF	404354.01	0750306.20	--	G	350.6	--	D	.3	32.33	--	--

Table 1. Location and well construction information for selected wells, Pohatcong Valley, Warren County, New Jersey.—Continued

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Well identification code	USGS well number	NJDEP permit number	USEPA well number	Local well number or name	Aquifer code	Latitude (NAD83)	Longitude (NAD83)	Lat/Long converted from NAD27 to NAD83	Method used to determine lat/long	Land surface altitude (feet above NAVD88)	Altitude converted from NGVD29 to NAVD88	Method used to determine altitude	Estimated accuracy of altitude (feet)	Depth of well (feet)	Top of open interval below land surface (feet)	Bottom of open interval below land surface (feet)
D404	410487	--	--	DOM D404	360KTTN	404411.40	0750239.74	--	G	345.8	--	D	.3	47.9	--	--
D406	410495	--	--	DOM D406	400PCMB	404550.79	0745705.51	--	G	503	--	M	10	271	--	--
D407	410496	--	--	DOM D407	400PCMB	404624.56	0745720.16	--	G	567	--	M	10	175	--	--
D409	410499	--	--	DOM D409	400PCMB	404645.92	0745820.31	--	G	493	--	M	10	365	--	--
D410	410501	--	--	DOM D410	360KTTN	404655.05	0745800.66	--	G	475	Yes	M	10	68.1	--	--
D411	410497	--	--	DOM D411	112SFDF	404627.61	0745951.37	--	G	426.2	--	D	.3	15.5	--	--
D412	410498	--	--	DOM D412	360KTTN	404629.14	0745952.65	--	G	438.0	--	D	.3	107	--	--
D413	410483	--	--	DOM D413	360KTTN	404349.35	0750336.40	--	G	395.4	--	D	.3	200	--	--
D414	410485	--	--	DOM D414	360KTTN	404354.28	0750337.97	--	G	401.2	--	D	.3	108.9	--	--
D415	410490	--	--	DOM D415	400PCMB	404458.88	0750203.66	--	G	475	--	M	10	85.35	60	85.35
D416	410491	--	--	DOM D416	400PCMB	404459.49	0750159.34	--	G	481	--	M	10	82	--	--
D433	410489	--	PV075DW	DOM D433	360KTTN	404414.86	0750329.49	--	G	469.6	--	D	.3	122.3	--	--
D437	410500	--	--	DOM D437	360KTTN	404652.17	0745819.09	--	G	458.1	--	D	.3	52	--	--
D443	410543	--	--	DOM D443	112SFDF	404656.31	0745746.58	--	G	509.7	--	L	.3	6	0	6
MW-42D	410416	24-33257	--	BRASS CASTLE ELEM - MW42D	371ALNN	--	--	--	G	437.8	--	L	.1	75	55	75
MW-42E	410413	24-33258	--	BRASS CASTLE ELEM - MW42E	371ALNN	--	--	--	G	436.8	--	L	.1	85	65	85
MW-42F	410415	24-33259	PVMWO05	BRASS CASTLE ELEM - MW42F	371ALNN	--	--	--	G	441.8	--	L	.1	85	65	85
MW-42H	410414	24-33261	PVMWO06	BRASS CASTLE ELEM - MW42H	371ALNN	--	--	--	G	438.1	--	L	.1	98	78	98
MW450	410410	24-21778-0	--	BASF MW450	112SFDF	404517.77	0745825.96	--	G	488.84	--	L	.1	38	18	38
MW451	410404	24-21776-3	--	BASF MW451	400PCMB	404518.97	0745823.01	--	G	494.06	--	L	.1	32.5	12.5	32.5
MW453	410481	--	--	MW453	360KTTN	404340.59	0750228.18	--	G	366.4	--	D	.3	95.57	--	--
MW454	410480	--	--	MW454	360KTTN	404325.31	0750240.22	--	G	387.4	--	D	.3	88.67	--	--
MW457	410546	24-21779-8	--	BASF MW-457	112SFDF	404517.56	0745823.77	--	G	497.55	--	L	.1	36.5	24.5	36.5
MW458	410492	--	--	MW458 MAXON MW-1	112SFDF	404538.22	0745819.24	--	G	459.6	--	D	.3	9.2	--	--

Table 1. Location and well construction information for selected wells, Pohatcong Valley, Warren County, New Jersey.—Continued

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MW459	410494	24-38677	--	MW459 MAXON MW-2	112SFDF	404540.31	0745820.00	--	G	465.3	--	D	.3	14	--	--
MW460	410493	24-38676	--	MW460 MAXON MW-3	112SFDF	404540.15	0745819.65	--	G	465.1	--	D	.3	22	--	--
MW465	410412	24-28101	--	WASH TWP MW465	112SFDF	404524.36	0750015.06	--	G	446.8	--	L	.1	24.5	9.5	24.5
MW469	410569	--	--	MW469 WBG	112SFDF	--	--	--	M	441	--	M	10	22.4	--	--
P483	410406	24-00491	--	AMERICAN CAN P483 INJECTION	374LSVL	404604.41	0745857.03	--	G	543.6	--	L	.1	245	93.08	245
P484	410335	24-04975	PVANC02	AMER CAN 1 (3)	374LSVL	404558	745858	Yes	M	542.3	--	L	.1	412	156.75	412
P485	410405	24-22618	PVANC01	AMERICAN CAN P485 (5)	374LSVL	404558.22	0745856.26	--	G	542.1	--	L	.1	383	140	383
P493	410381	24-03250	--	NJ BELL TELE-PHONE CO.	371ALNN	404528	745852	Yes	M	469	Yes	M	10	154	32	154
P494	410346	24-22566	--	BASF 2 (REPL-1) P494	374LSVL	404522	745819	Yes	M	494	--	M	10	496	74	496
P495	410411	24-31397	--	BASF 3 P495	400PCMB	404520.44	0745821.14	--	G	497	--	M	10	599	103	599
PVANC03	410559	24-26014	PVANC03	PVANC03	374LSVL	404603.36	0745857.07	--	G	540.72	--	L	.3	129	119	129
PVANC05	410560	24-26015	PVANC05	PVANC05	374LSVL	404604.19	0745851.78	--	G	546.0	--	L	.3	131	121	131
PVANC06	410553	24-25965	PVANC06	PVANC06	374LSVL	404602.64	0745850.12	--	G	546.28	--	L	.3	136	126	136
PVANC29	410552	24-28992	PVANC29	PVANC29	374LSVL	404559.11	0745858.08	--	G	514.53	--	L	.1	101	81	101
PVANC30	410551	24-28995	PVANC30	PVANC30	374LSVL	404559.07	0745858.22	--	G	514.43	--	L	.1	245	220	245
PVBAS01	410505	24-40824	PVBAS01	PVBAS01	400PCMB	404519.69	0745824.37	--	G	493.20	--	L	.1	210	190	210
PVBAS02	410504	24-40825	PVBAS02	PVBAS02	112SFDF	404519.64	0745824.49	--	G	493.48	--	L	.1	80	70	80
PVBAS03	410507	24-40826	PVBAS03	PVBAS03	400PCMB	404521.03	0745819.22	--	G	498.93	--	L	.1	80	70	80
PVBMC01	410527	24-40832	PVBMC01	PVBMC01	371ALNN	404454.58	0750045.60	--	G	381.42	--	L	.1	92	72	92
PVBMC02	410526	24-40833	PVBMC02	PVBMC02	112SFDF	404454.45	0750045.54	--	G	380.78	--	L	.1	43	33	43
PVCHU02	410474	24-38708	PVCHU02	PVCHU02	371ALNN	404528.08	0750007.99	--	G	425.71	--	L	.01	129	104	129
PVFCC12	410472	24-38706	PVFCC12	PVFCC12-MW1	371ALNN	404522.89	0745953.73	--	G	393.71	--	L	.01	155	130	155
PVFCC13	410473	24-38705	PVFCC13	PVFCC13	371ALNN	404524.22	0745954.42	--	G	392.74	--	L	.01	67.5	57.5	67.5
PVGPU01	410471	24-38717	PVGPU01	PVGPU01-MW1	374LSVL	404601.38	0745920.86	--	G	514.54	--	L	.01	185	160	185

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PVGSS04	410475	24-38723	PVGSS04	PVGSS04-MW1	374LSVL	404544.78	0745744.71	--	G	501.71	--	L	.01	77	67	77
PVGSS05	410476	24-38722	PVGSS05	PVGSS05-MW1	374LSVL	404545.1	0745744.74	--	G	500.58	--	L	.01	123	98	123
PVHBR07	410428	24-32678	PVHBR07	PVHBR07	112SFDF	404515.55	0745846.7	--	G	472.11	--	L	.01	38	23	38
PVHBR08	410545	--	PVHBR08	PVHBR08	112SFDF	404515.62	0745845.55	--	G	476.57	--	L	.1	71	61	71
PVLNL05	410512	24-40805	PVLNL05	PVLNL05	374LSVL	404531.92	0745843.16	--	G	463.42	--	L	.1	145	126	146
PVLNL06	410513	24-40806	PVLNL06	PVLNL06	112SFDF	404532.04	0745843.21	--	G	463.65	--	L	.1	87	77	87
PVLNL07	410511	24-40807	PVLNL07	PVLNL07	112SFDF	404531.83	0745843.11	--	G	463.27	--	L	.1	72.3	62.3	72.3
PVMSW01	410282	24-08261	PVMSW01	WASHINGTON 1972	374LSVL	404527	745836	Yes	M	440.6	--	L	.3	345	88	345
PVMSW02	410021	24-16653	PVMSW02	WASHINGTON 5	360KTTN	404519	745735	Yes	M	459	Yes	M	10	--	152	407
PVMSW04	410316	24-12183	PVMSW04	DALE AVE 4	360KTTN	404530	750006	Yes	M	417	--	L	.3	184	143	184
PVMVS06	410514	24-40808	PVMVS06	PVMVS06	112SFDF	404533.93	0745844.15	--	G	469.19	--	L	.1	81	71	81
PVMWO01	410424	24-28052	PVMWO01	PVMWO01	360KTTN	404246.84	0750330.08	--	G	419	Yes	M	10	81	61	81
PVMWO02	410429	24-30551	PVMWO02	PVMWO02	374LSVL	404546.48	0745735.75	--	G	504	Yes	M	10	77	62	77
PVMWO03	410425	24-28050	PVMWO03	PVMWO03	360KTTN	404248.17	0750321.28	--	G	399	Yes	M	10	74	54	74
PVMWO04	410430	24-30550	PVMWO04	PVMWO04	374LSVL	404546	0745736	--	G	504	Yes	M	10	100	85	100
PVRAB04	410478	24-38711	PVRAB04	PVRAB04-MW1	374LSVL	404626.14	0745845.98	--	G	510.24	--	L	.01	121	111	121
PVRAB05	410477	24-38710	PVRAB05	PVRAB05-MW1	374LSVL	404626.04	0745846.09	--	G	510.76	--	L	.01	174	154	174
PVSMC01	410544	--	PVSMC01	PVSMC01	112SFDF	404514.47	0745832.73	--	G	486.73	--	L	.1	19	--	--
PVSMC02	410503	24-40816	PVSMC02	PVSMC02	112SFDF	404512.61	0745832.43	--	G	494.23	--	L	.1	95.5	75.5	95.5
PVSMC03	410502	24-40817	PVSMC03	PVSMC03	112SFDF	404512.58	0745832.26	--	G	494.21	--	L	.1	31	21	31
PVSTJ01	410515	24-40803	PVSTJ01	PVSTJ01	371ALNN	404544.36	0745932.09	--	G	469.44	--	L	.1	125.5	105.5	125.5
PVSTJ01A	410516	24-41076	PVSTJ01A	PVSTJ01-A	374LSVL	404544.68	0745932.45	--	G	470.01	--	L	.1	260	240	260
PVTVN13	410509	24-40812	PVTVN13	PVTVN13	374LSVL	404521.35	0745829.27	--	G	478.30	--	L	.1	252	232	252
PVTVN14	410508	24-40813	PVTVN14	PVTVN14	112SFDF	404521.20	0745829.30	--	G	478.13	--	L	.1	85.5	75.5	85.5
PVTVN15	410506	24-40814	PVTVN15	PVTVN15	112SFDF	404519.82	0745827.58	--	G	490.25	--	L	.1	71	61	71
PVTVN16	410510	24-40815	PVTVN16	PVTVN16	400PCMB	404522.39	0745822.93	--	G	485.13	--	L	.1	100	90	100
PVVAN01	410528	24-40820	PVVAN01	PVVAN01	374LSVL	404605.85	0745851.70	--	G	558.91	--	L	.1	213	193	213

Table 1. Location and well construction information for selected wells, Pohatcong Valley, Warren County, New Jersey.—Continued

[USGS, U.S. Geological Survey; NJDEP, New Jersey Department of Environmental Protection; USEPA, U.S. Environmental Protection Agency; 112SFDF, Stratified drift; 360KTTN, Kittatiny Group, 371ALNN, Allentown Formation; 374LSVL, Leithsville Formation; 400PCMB, Precambrian gneiss; lat, latitude; long, longitude; NAD27, North American Datum of 1927; NAD83, North American Datum of 1983; NGVD29, National Geodetic Vertical Datum of 1929; NAVD88, North American Vertical Datum of 1988; M, map; G, Global Positioning System; L, Leveled; D, Differential Global Positioning System; --, no data or not applicable]

Well identification code	USGS well number	NJDEP permit number	USEPA well number	Local well number or name	Aquifer code	Latitude (NAD83)	Longitude (NAD83)	Lat/Long converted from NAD27 to NAD83	Method used to determine lat/long	Land surface altitude (feet above NAVD88)	Altitude converted from NGVD29 to NAVD88	Method used to determine altitude	Estimated accuracy of altitude (feet)	Depth of well (feet)	Top of open interval below land surface (feet)	Bottom of open interval below land surface (feet)
PVVAN02	410529	24-40821	PVVAN02	PVVAN02	112SFDF	404606.00	0745851.74	--	G	559.19	--	L	.1	145	135	145
PVVCA05	410536	24-28488	PVVCA05	PVVCA05	360KTTN	404234.27	0750401.52	--	G	350.73	--	L	.1	60	22	60
PVVCA06	410535	24-30571	PVVCA06	PVVCA06	360KTTN	404230.92	0750408.08	--	G	342	--	L	.1	80	58	80
PVVTC07	410469	24-39141	PVVTC07	PVVTC07-MW1	374LSVL	404549.53	0745853.97	--	G	501.40	--	L	.01	15	5	15
PVVTC13	410517	24-40800	PVVTC13	PVVTC13	374LSVL	404549.46	0745854.12	--	G	500.75	--	L	.1	156	136	156
PVVTC14	410518	24-40801	PVVTC14	PVVTC14	112SFDF	404549.52	0745853.88	--	G	501.88	--	L	.1	93	83	93
PVVTC15	410520	24-40802	PVVTC15	PVVTC15	112SFDF	404550.57	0745858.02	--	G	501.37	--	L	.1	133	123	133
PVWBG01	410547	24-27713	PVWBG01	PVWBG01	112SFDF	404527.86	0745838.49	--	G	443.3	--	L	.3	60	50	60
PVWBG02	410549	24-27132	PVWBG02	PVWBG02	112SFDF	404528.26	0745838.17	--	G	442.77	--	L	.3	20	5	20
PVWBG03	410548	24-27131	PVWBG03	PVWBG03	112SFDF	404527.90	0745840.08	--	G	444.4	--	L	.3	20	5	20
PVWBG04	410550	24-27133	PVWBG04	PVWBG04	112SFDF	404529.16	0745839.64	--	G	446.32	--	L	.3	23	8	23
PVWBG11	410467	24-38701	PVWBG11	PVWBG11-MW1	374LSVL	404526.89	0745839.36	--	G	442.77	--	L	.01	116	106	116
PVWBG12	410468	24-38700	PVWBG12	PVWBG12-MW1	374LSVL	404527.10	0745839.46	--	G	442.86	--	L	.01	169	149	169
PVWCC14	410462	24-38696	PVWCC14	PVWCC14-MW1	371ALNN	404413.27	0750226.66	--	G	343.21	--	L	.01	42	32	42
PVWCC15	410463	24-38697	PVWCC15	PVWCC15-MW1	371ALNN	404413.27	0750226.95	--	G	343.68	--	L	.01	96	71	96
PVWCV05	410464	24-39495	PVWCV05	PVWCV05-MW1	371ALNN	404426.95	0750130.03	--	G	404.88	--	L	.01	145	115	145
PVWLY07	410470	24-38729	PVWLY07	PVWLY07-MW1	374LSVL	404551.22	0745859.55	--	G	500.95	--	L	.01	250	225	250
PVWLY08	410524	24-40818	PVWLY08	PVWLY08	374LSVL	404554.55	0745856.40	--	G	517.52	--	L	.1	242	222	242
PVWLY09	410523	24-40819	PVWLY09	PVWLY09	112SFDF	404554.42	0745856.31	--	G	517.61	--	L	.1	123	113	123
PVWLY12	410525	24-41617	PVWLY12	PVWLY12	374LSVL	404554.55	0745856.40	--	G	517.52	--	L	.1	151	136	151
PVWPC13	410465	24-38691	PVWPC13	PVWPC13-MW1	374LSVL	404513.53	0745907.33	--	G	455.30	--	L	.01	45	35	45
PVWPC14	410466	24-38692	PVWPC14	PVWPC14-MW1	374LSVL	404513.86	0745907.48	--	G	453.96	--	L	.01	237	217	237